

Water fluxes on different spatial and temporal scales in a semi-arid steppe environment: experimental and modelling approaches

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1 Extended summary

1.1 Introduction

Water is a key element linking the ecological processes in the soil, vegetation and atmosphere. Particularly in arid and semiarid environments water presents a limiting factor in the biogeochemical cycle. To name only a few, nutrient turnover, biomass production, gas exchange between soil, vegetation and atmosphere or runoff generation and coupled matter transport are closely dependent on the presence of water. Soil water has a significant function controlling these processes. However, the quantity and distribution of soil water content is affected by land cover, surface and soil properties and atmospheric circulation patterns. According to Rodriguez-Iturbe et al. (2001), climate and soil characteristics act externally, while vegetation characteristics are closely linked to soil moisture dynamics. The processes act at different spatial and temporal scales. Nevertheless, soil water storage at the interface between soil, water and atmosphere is affected and in consequence, will influence land use decisions.

The heterogeneous distribution of physical properties of the biogeosphere influences hydrological fluxes, e.g. infiltration or surface runoff. Hence, even under comparable precipitation conditions, soil moisture patterns with a varying share to runoff generation evolve. The distribution of soil moisture patterns is related to the heterogeneity of biogeospheric properties and atmospheric processes. In pristine catchments, soil moisture distribution and storage is controlled by the prevailing environmental conditions, whereas in areas under human impact, land use management will alter the factors controlling soil moisture patterns and runoff generation (Dunn and Mackay, 1995; Doe et al., 1996; Bormann et al., 1999; Hernandez et al., 2000). For example, Li et al. (2000) and Golodets and Boeken (2006), showed in plot and field scale studies that grazing affects micrometeorological fluxes and soil moisture storage.

From a macroscale perspective (i.e. $> 1000 \text{ km}^2$), atmospheric circulation patterns control soil moisture dynamics (Vinnikov et al., 1996; Entin et al., 2000). When zooming in to smaller scales (i.e. catchment, hillslope or plot scale), topography and physical properties of soils and vegetation become more relevant for soil moisture variability. Western et al. (1999) refer to an *organized* soil moisture distribution when it correlates to catchment characteristics, i.e. topographic indices. This is particularly the case under wet conditions, while under more dry conditions, random soil moisture patterns evolve that are disconnected from topographic catchment characteristics. In addition to spatial controls, there is a temporal scale altering processes of soil moisture distribution. For example, vegetation dynamics cause seasonal variations in plant water demand and hence, the spatial patterns of soil moisture may change in the course of vegetation cycles.

Identifying the factors controlling spatial variability of soil moisture is crucial for understanding runoff generation. On the one hand, surface and subsurface characteristics influence soil water storage. Depending on catchment or hillslope specific precipitation thresholds, soil moisture patterns connect to each other (Tromp-van Meerveld and McDonnell, 2006). Particularly in semi-arid areas that are characterized by seasonality of precipitation in combination with high evapotranspiration rates, areas with higher connectivity may be the only sources for runoff generation. Thus, information about soil moisture variability in the context of landscape properties is required when runoff generation and coupled matter fluxes

have to be modelled in a process based way. On the other hand, atmospheric forcing controls the direction of water fluxes. When the warm season with highest energy input coincides with the wet season, plant water demand, atmospheric water deficit and high temperatures will result in high evapotranspiration rates. In warm arid and semi-arid climates, potential evapotranspiration exceeds precipitation by far. In contrast to humid zones, evapotranspiration in arid and semi-arid areas has a strong influence on hydrological fluxes and hence, needs to be implemented in model applications in an adequate way. As field measurements are difficult, evapotranspiration rates are usually derived from climate and vegetation parameters, limiting accuracy of the estimates with respect to data quality and availability.

Although soil moisture is a key variable for the processes previously discussed, ubiquitous information often is lacking as ground based measurements of soil moisture are labour-intensive, especially when they have to be synchronised over large areas. Plot and field scale measurements usually cover only a limited area and time and hence, need to be supported by model studies (e.g. for comparison of land use scenarios or upscaling purposes). Improving our understanding of soil moisture dynamics in a landscape context is a prerequisite when generalization and upscaling rules for model approaches are required. During the last years, research focused on the effects of substrate, land cover and land use on soil moisture storage (Rodríguez-Iturbe and Porporato, 2004), as well as on spatial and temporal dimensions of soil moisture variability (Choi et al., 2007; D'Odorico et al., 2007). The results are often ambiguous. In a comprehensive study, Famiglietti et al. (2008) show that the relationship between soil water dynamics and the spatial scale considered is not linear. Still, open questions remain and require both experimental and modelling work.

In-situ measurements of soil moisture on the field or basin scale are labour intensive and hence, the sampling volume and temporal resolution is restricted. Nevertheless, soil moisture information is also required on larger scales, which requires alternative sampling approaches. Remote sensing products offer soil moisture information on spatial scales where ground-based measurements are not feasible. Several studies have tested the potential of airborne and spaceborne techniques to assess soil moisture (Sano et al., 1998; Cosh et al., 2004; Bindlish et al., 2006). Particularly active radar techniques are promising as they are not affected by cloud cover like optical systems. Yet, the accuracy of soil moisture data derived from airborne and spaceborne platforms varies as surface properties, i.e. roughness length, largely affect the backscatter signal. In consequence, careful calibration of remote sensing products with ground-based measurements is necessary to obtain reliable soil moisture information. The question arises if the quality of remote sensing products matches the requirements of hydrological studies, and how the effort of ground-based measurements can be reduced for frequent comparisons with remote sensing data while at the same time uncertainty about spatial variability of soil moisture will remain on an acceptably low level.

The discussion shows that understanding soil moisture dynamics is essential for experimental and numerical ecohydrological applications from micro- to macroscale. Despite long-term research, open questions on the effect of land use on spatial and temporal soil moisture dynamics remain. Also, generalizing in-situ measurements for applications on larger scales is still subject to considerable uncertainties. The aim of the presented project is to provide information to better understand key hydrological processes in a semi-arid steppe environment.

Based on field and modelling studies, the influence of land use, i.e. grazing, on the spatial and temporal dynamics of water fluxes was analysed. The study was conducted within the research group 536 „Matter fluxes in grasslands of Inner Mongolia as influenced by stocking rate (MAGIM)“ which is funded by the German Science Foundation (DFG). The results shall finally provide information to meet the demands for landscape scale ecohydrological modelling of an ungauged, semi-arid watershed. In the context of the research framework, the central questions of this study are as follows:

- 1) How do land cover properties, and hence, land use, influence soil moisture storage? Can we quantify the influence of grazing on soil moisture, or does atmospheric forcing overlay all other factors?
- 2) Can we quantify the role of evapotranspiration with model approaches to yield acceptable results for discharge calculations, and which model is suited best for the study environment?
- 3) How much measurements effort is required to accurately assess soil moisture dynamics and to which extent can remote sensing products estimate soil moisture storage in order to provide catchment scale hydrological information in data sparse regions?

The questions raised shall increase knowledge on the effect of land use on hydrology in semi-arid continental catchments on the one hand, and provide methods to assess these effects on different scales on the other hand. Particularly the role of soil moisture as a key component for runoff generation needs to be clarified, as steppe environments carry important environmental and socioeconomic ecosystem functions, and hence their importance expands far beyond the studied scale.

1.2 Study area

The study was carried out in the Xilin river catchment in the Autonomous Region Inner Mongolia (PR China). It is part of the Eurasian steppe belt. The entire catchment covers an area of roughly 10'000 km², but the area delineated for this study is terminated by the gauging station south of the city of Xilinhote and covers 3600 km² (Fig. 1.1). The Xilin river drains into an endorheic basin. It originates in the Daxingan mountains in the south-eastern part of the catchment, crosses a sand dune belt and flows towards the city of Xilinhote to finally run dry in a depression north of the catchment.

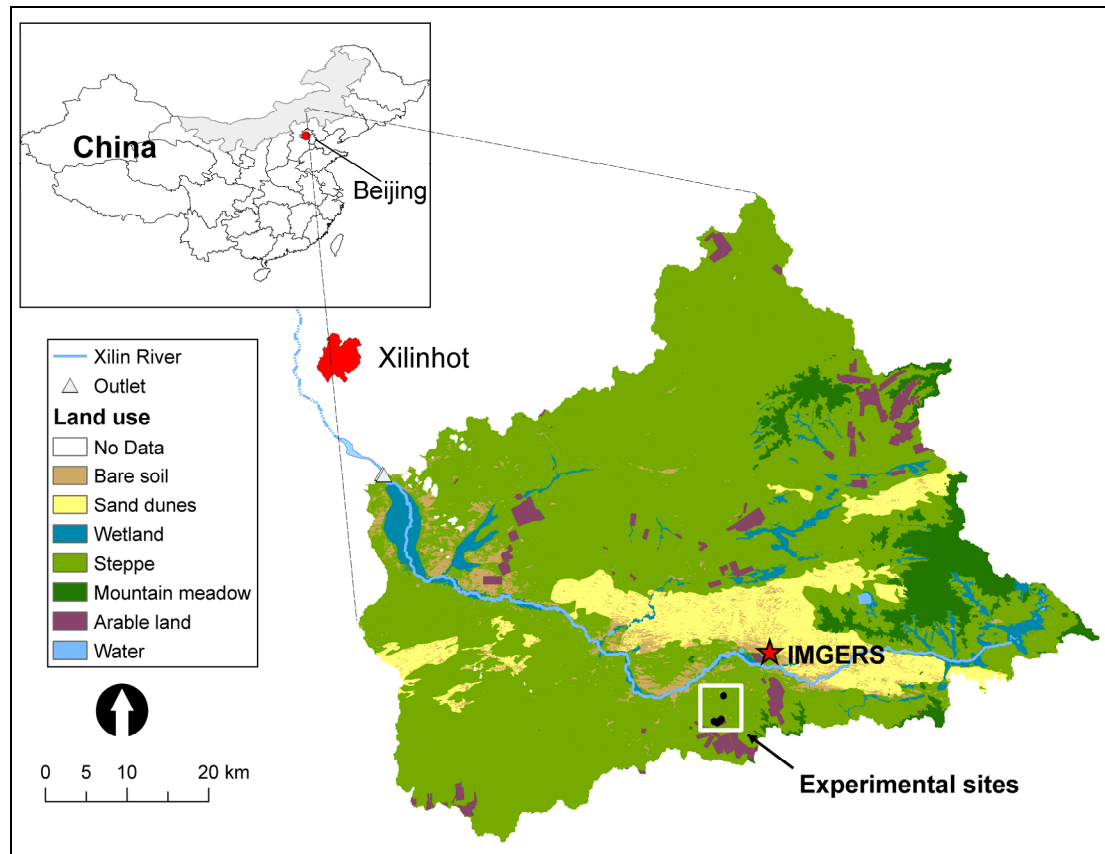


Figure 1.1. Location and land use of the Xilin river catchment.

1.2.1 Climate

The climate in the Xilin river catchment is continental with cold, dry winters and warm, wet summers (Fig. 1.2). Annual average temperature is -2.3°C , with mean maximum and minimum ranging from $+18^{\circ}\text{C}$ in summer to -23°C in winter. From 1982 to 2006, mean annual precipitation was 334 mm at the Inner Mongolian Grassland Ecosystems Research Station IMGERS (data provided by IMGERS) and 278 mm in Xilinhot (data provided by the Hydrometeorological Office Xilinhot), but interannual variations are high (Chen, 1988). The gradient between the IMGERS and the city of Xilinhot located 65 km to the northwest results from increasing continentality along the track of the East Asian monsoon.

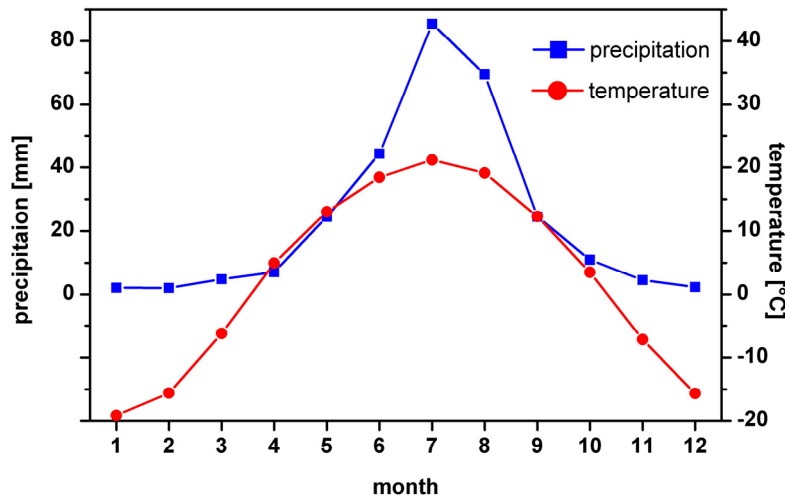


Figure 1.2. Mean precipitation and temperature (1957-2003), Xilinhot.

1.2.2 Vegetation and land cover

Potential natural vegetation is *Stipa grandis* and *Leymus chinensis* steppe. Yet, overgrazing leads to several stages of degradation and a shift towards *Cleistogenes squarrosa*-*Artemisia frigida* or *Artemisia frigida* communities (Tong et al., 2004). Within the sand dune belt, Siberian Elm (*Ulmus pumilla*) occurs; outside the sand dunes, trees only grow in sheltered trenches or in artificial plantations along the settlements. Traditional use of the steppe is nomadic pastoralism, but animal husbandry in settlements without rotating the pasture has become much more important all over Inner Mongolia throughout the last decades (Williams, 2002). Crop land covers only a small portion of the catchment area as climate conditions do not favour intensive crop production.

1.2.3 Geology, geomorphology and soils

Information on the geology of the Xilin catchment is incomplete, but four main lithologic units can be derived: (1) volcanic rocks, mainly plateau basalts in the southwest, (2) quaternary deposits in the middle and eastern parts, (3) shales in the north and (4) igneous rocks along a southwest-northeast stretch between the latter two. The relief is undulating with a mean slope of 2.7°. Elevation ranges from approx. 1600 m a.s.l. in the Daxingan Mountains in the east to 1000 m a.s.l. at the catchment outlet in the northwest. Main soil types are calcic chernozems and kastanozems; saline soils (solonchaks) occur in the northwestern part of the catchment. In the sand dune belt, soils are less developed with low humus content and low aggregate stability. Soil properties based on the studies of Steffens et al. (2008). and Hoffmann et al. (2008) are given in chapter 3.2.1.

1.2.4 Runoff generation

Long term discharge observations exist for the gauging station south of Xilinhot from 1957-2004 (Fig. 1.3). Mean discharge of the Xilin river is $0.6 \text{ m}^3/\text{s}$, but the annual average does not reflect the seasonality of runoff generation in this semi-arid environment. Despite the low precipitation during winter, peak discharge occurs during snow melt in spring (March - May), whereas the summer precipitation maximum is only reflected by a secondary discharge peak. The discharge coefficient k (constituted by the ratio of mean monthly to mean annual discharge) is 4.1 in April, and only 1.3 in July which, in general, receives most precipitation. During winter the Xilin river is frozen. Therefore, in general no discharge is observed from December until February.

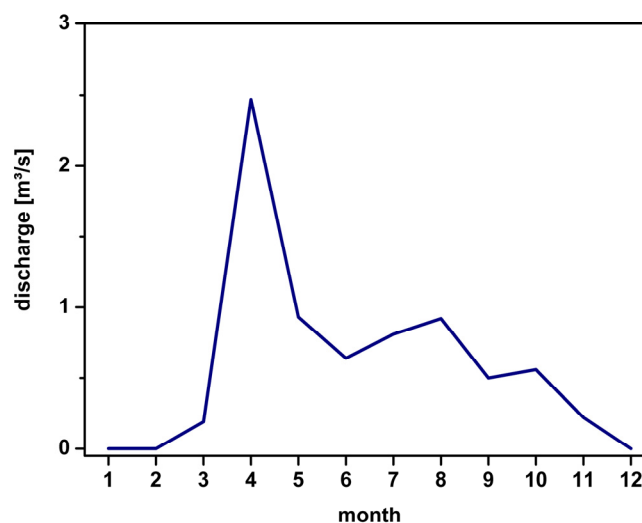


Figure 1.3. Mean discharge (1957-2003), Xilinhot.

1.2.5 Field scale experimental sites

Soil moisture measurements were taken on five sites with different grazing intensities located 8 km south of IMGERS. The sites comprise two enclosures which have been fenced and protected from grazing since 1979 (ug79) and 1999 (ug99). The three grazed sites represent (1) winter grazing (wg) with a grazing intensity of 1.5 sheep/ha, (2) continuous grazing (cg) with a grazing intensity of 2 sheep/ha, and (3) heavy grazing (hg) with a grazing intensity of 4 sheep/ha. On each site a measurement grid covering approximately 1.5 ha (105 m x 135 m) with 100 sampling points was installed. The regular spacing was 15 m; the spacing of five embedded nests was 5 m (see chapter 3.2.2, Fig. 3.2). The effect of various grazing intensities is reflected in the vegetation characteristics of the sites. Vegetation cover, i.e. leaf area index (LAI) and biomass production decrease with increasing grazing intensity (Fan et al., 2008; Gao et al., 2008). Likewise, litter cover decreases with increasing grazing intensity, leading to a higher share of bare soil on the grazed sites. Due to the different duration of enclosure, the ug79 site has a higher degree of shrub vegetation than the ug99 site.

1.3 On the role of evapotranspiration: comparing evapotranspiration methods with field data

1.3.1 Introduction

Evapotranspiration is a major component of hydrological fluxes in warm arid and semi-arid environments. In the study area, potential evapotranspiration (PET) (measured with a class A pan at IMGERS) in 2004, 2005 and 2006 was 1700 mm in average, exceeding precipitation rates by far. Hence, evapotranspiration is a limiting factor for plant growth and runoff generation. Quantifying the amount of evapotranspiration therefore is essential in order to optimize hydrological models. After hydrological simulations with the Soil and Water Assessment Tool (SWAT, (Arnold et al., 1998)) proved to show rather poor results for the Xilin river catchment, the effect of different evapotranspiration methods on discharge was evaluated. The comparison showed the reliability and applicability of simple versus complex evapotranspiration methods, and whether a method could be identified that fits best to simulate hydrology in the semi-arid steppe environment of the Xilin catchment. Four evapotranspiration methods with varying data requirements were chosen to test the effect of different degrees of complexity on hydrological simulations: (1) Hargreaves, (2) Makkink, (3) Priestley-Taylor, (4) Penman-Monteith. The first one is the most simple and is based on temperature solely while the latter is the most complex. Evapotranspiration following the Makkink method was calculated externally and read into the SWAT model. The other three methods are already implemented in SWAT and could directly be used within the hydrological model. The data requirements for the models are given in chapter 2.2.2 (Table 2.1).

The evapotranspiration methods were run within the SWAT routine for the vegetation period 2004 and 2005. The results were compared with evapotranspiration rates derived from eddy covariance measurements in the study area. While precipitation during the vegetation period 2004 was 288 mm, it was much less in 2005 (125 mm). Hence, the performance of the evapotranspiration methods in an average and in a dry year was compared.

1.3.2 Results and discussion

The results show that all evapotranspiration methods follow the seasonal course of the measured evapotranspiration. Yet, the methods either overestimate (Hargreaves and Makkink) or underestimate (Priestley-Taylor and Penman-Monteith) evapotranspiration measured with the eddy-covariance system. Figure 1.4 shows daily modelled and measured evapotranspiration for the study periods in 2004 and 2005. Although the study period in 2004 is shorter than in 2005, evapotranspiration rates are generally lower in the dryer year 2005, indicating that rainfall is the main source of evapotranspiration, and that groundwater contribution (i.e. through capillary rise) to evapotranspiration during periods of water scarcity can be neglected. This was confirmed by Wen et al. (2007) who showed that almost all precipitation is consumed by evapotranspiration in the study area. The results for the data intensive Priestley-Taylor and Penman-Monteith methods show higher amplitudes compared to measured data and to the simpler Hargreaves and Makkink methods.

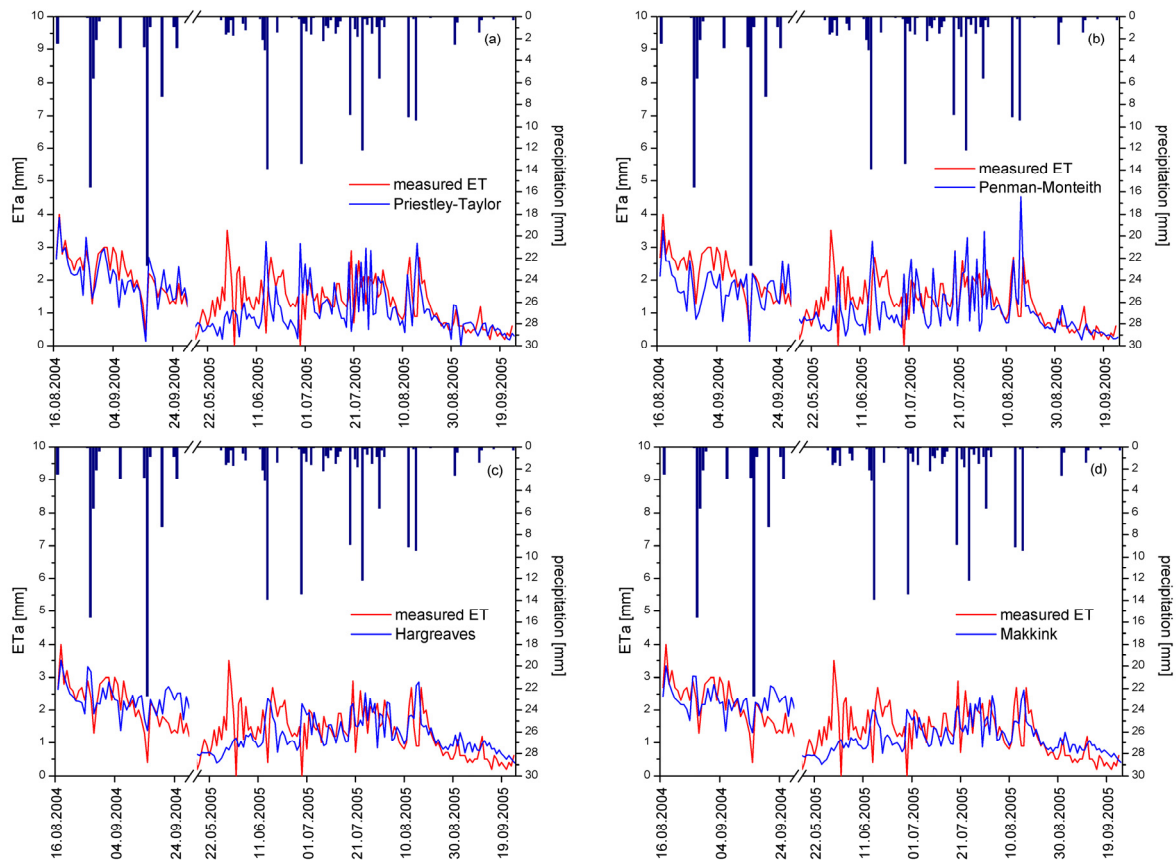


Figure 1.4. Observed vs. measured actual evapotranspiration (ETa): (a) Priestley-Taylor, (b) Penman-Monteith, (c) Hargreaves, (d) Makkink (Schneider et al., 2007, see chapter 2.3).

The correspondence between calculated and measured data as given by the Nash-Sutcliffe-Efficiency is negative for all methods and lowest for the Priestley-Taylor and Penman-Monteith method. Under average precipitation conditions in 2004 the deviation from measured evapotranspiration remains moderate; however, it increases in the dryer study period of 2005. Particularly the deviation from measured evapotranspiration (Table 1.1) is highest for the Priestley-Taylor and Penman-Monteith methods.

Table 1.1. Summary of precipitation and observed and calculated ET for 2004 and 2005 (Schneider et al. 2007, see chapter 2.3).

	Rainfall	Observed ET	PT	PM	HG	MK
16.8. – 30.9.2004	82.6	99.3	92.1	81.3	107.4	104.3
Deviation from observed (%)			-7	-18	8	5
15.5. – 24.9.2005	113.1	174.5	140.0	145.0	164.0	155.3
Deviation from observed (%)			-19.8	-16.9	-6	-11.2

While the evapotranspiration methods show differences in their performance, their effect on improving discharge simulations is minor. The Hargreaves and Makkink methods prove to give best results, but even when applying them in the hydrological model, discharge simulations remain poor. With any of the discussed evapotranspiration methods, the SWAT model results are biased from observed discharge data. The model calculates 15%-30% of actually observed discharge only. As calculated and measured evapotranspiration rates during the study period show much less differences, replacing an evapotranspiration method for another will not account for major corrections of hydrological simulations.

1.3.3 Conclusions

Although the methods calculate different evapotranspiration rates, the variation remains moderate compared to the huge bias between observed and modelled discharge. Drier conditions promote deviations more pronounced than average conditions. Yet, all evapotranspiration methods match the course of the observed evapotranspiration. Firstly, this suggests that the impact of a particular evapotranspiration method on discharge calculations can be neglected in the study environment. The question remains whether this holds true for hydrological simulations in catchments with different climate conditions, e.g. when evapotranspiration is not dominating hydrological processes and hence may become a more sensitive model component. Secondly, other factors than evapotranspiration obviously cause the huge bias between discharge observations and simulations. Steady groundwater inflow which is not captured by the model structure or the available input data might be an explanation for these differences. The very simple Hargreaves method proved to perform best under the climatic conditions in the study environment. In data sparse regions like the Xilin river catchment, access to and availability of climate data is restricted, so the Hargreaves method offers a powerful tool for feasible evapotranspiration estimates. Nevertheless, other methods might outperform the Hargreaves method under different environmental conditions.

1.4 Effects of grazing on spatial and temporal dynamics of soil water

1.4.1 Background

Evapotranspiration proved to be a major component of the hydrological cycle in the study area. On a local scale (i.e. plot and field scale) with invariant climatic and topographic boundary conditions, evapotranspiration patterns and the partitioning into its components will be affected by vegetation and soil characteristics. The predominant land use in the study area is sheep grazing. Biomass production, plant composition and plant cover on intensively grazed pastures are distinctively different from ungrazed areas or those areas with lower grazing intensity (Tong et al., 2004; Gao et al., 2008).

To assess the impact of different grazing intensities on soil moisture, measurements were taken on the five experimental sites ug79, ug99, wg, cg and hg (for detailed characterization see chapter 1.2 and 3.2.2). FDR (frequency domain

reflector) measurements in the top 0.06 m of the soil were timed according to precipitation and drying cycles so that soil moisture dynamics on the sites could be observed for various stages of wetness and dryness. A wetting-drying cycle with daily measurements after a precipitation event of 16 mm was captured in June 2006. In total, 500 measurements (i.e. 100 measurements per site) were taken per sampling date. With this sampling design, spatial and temporal dynamics of soil moisture can be assessed, and the effect of different surface characteristics (i.e. plant cover and plant composition) on soil moisture can be analysed. The high number of sampling points single out artefacts originating from small scale heterogeneity of the factors determining soil moisture storage, and provide a more comprehensive understanding than accessible via point measurements.

Based on a geostatistical analysis, the similarity of spatially distributed sampling points can be acquired. The semivariance is a measure for the similarity (or dissimilarity) of data pairs with respect to their sampling separation. In general, semivariance increases with increasing sampling separation, indicating higher dissimilarity between data pairs (Olea, 1999). The variogram function is fitted to the semivariance $\gamma(h)$ of N data pairs $z(x_i)$ depending on their sampling separation h .

$$\gamma(h) = \frac{1}{2N} \sum \{z(x_i) - z(x_i + h)\}^2 \quad (1)$$

The shape of the variogram function is determined by the nugget (i.e. noise due to small scale variability below the sampling resolution and precision of the FDR device), the sill (i.e. overall variability of the measurements), and the range (i.e. sampling separation up to which correlation exists). Based on the variogram function, the soil measurements were interpolated with kriging.

1.4.2 Temporal dynamics of soil moisture as subject to grazing intensity

After a dry period, a precipitation event on three consecutive days with 16 mm in total raised mean soil moisture on the experimental sites to 0.20-0.25 m³/m³ (Fig. 1.5, Table 1.2). Soil moisture prior to the precipitation event was very low (< 0.1 m³/m³). Although the experimental sites are close to each other, and hence, precipitation values are expected to correspond, the initial wetting did not result in uniform mean soil moisture values. The ug99 has highest and the hg site has lowest mean soil moisture throughout the monitoring period. The daily measurements following the precipitation events showed that all sites dried steadily. Nevertheless, from a very similar mean moisture status, the wg and cg sites show a higher drying rate than the ug99 site. The two ungrazed sites, ug99 and ug79, are adjacent, but protection from grazing does not necessarily result in similar mean field soil moisture. As Table 1.2 shows, the two sites are significantly different from each other on all sampling dates. Soil texture on the two sites is comparable (Steffens et al., 2008), and hence, the differences are assigned to the vegetation properties: the ug79 site bears more shrub-like vegetation with a higher leave area than the ug99 site with more grass species. This might lead to higher interception of precipitation and less infiltration on the ug79 site. In consequence, a linear rating of grazing impact on soil moisture is not possible as (1) mean soil moisture on both, the ug79 site and the hg site are

significantly different from the ug99 site and (2) mean soil moisture on the moderately grazed wg and cg sites is higher than on the ug79 site.

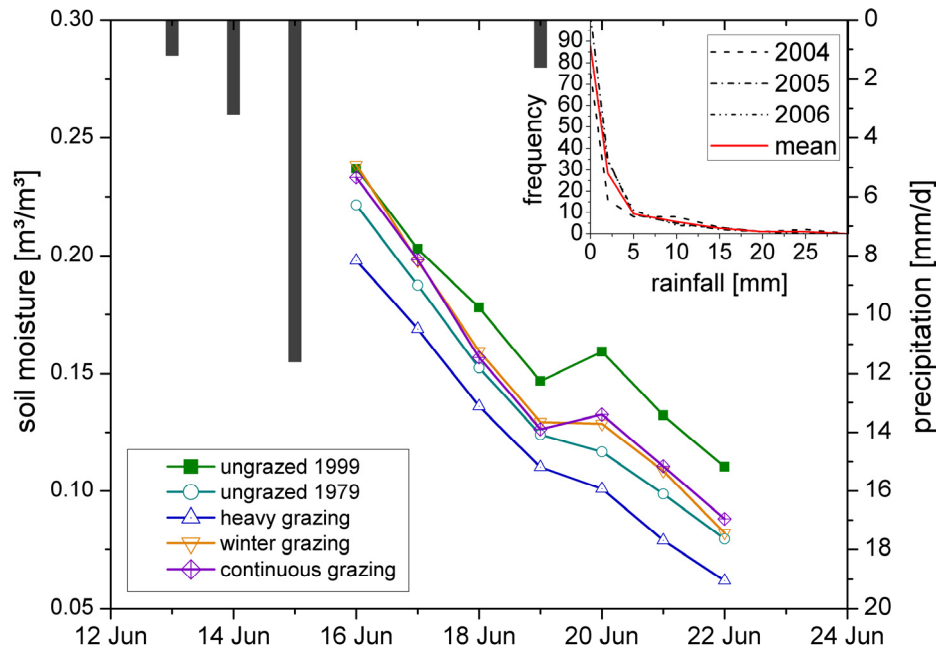


Figure 1.5. Mean ($n=100$) soil moisture and daily sums of precipitation during the sampling period in 2005. Inlet shows frequency distribution of rainfall events for 2004, 2005, 2006 and the 3-year average (Schneider et al. 2008, see chapter 3.3.1).

Table 1.2. Mean and standard deviation of soil moisture measurements on the five grazing sites (Schneider et al. 2008, see chapter 3.3.1).

Date	ungrazed 1999		ungrazed 1979		winter grazing		continuous grazing		heavy grazing	
	mean	SD	mean	SD	mean	SD	mean	SD	mean	SD
08.06.2005*	0.046 ^a	0.018	0.073 ^b	0.017	0.041 ^{cd}	0.011	0.046 ^{ad}	0.015	0.043 ^{ad}	0.011
16.06.2005	0.237 ^a	0.020	0.221 ^b	0.027	0.239 ^a	0.015	0.233 ^a	0.019	0.198 ^c	0.024
17.06.2005	0.203 ^a	0.021	0.187 ^b	0.025	0.198 ^a	0.018	0.198 ^a	0.019	0.169 ^c	0.021
18.06.2005	0.178 ^a	0.021	0.152 ^b	0.024	0.160 ^b	0.015	0.157 ^b	0.018	0.136 ^c	0.019
19.06.2005	0.147 ^a	0.021	0.124 ^b	0.026	0.129 ^b	0.013	0.126 ^b	0.017	0.110 ^c	0.018
20.06.2005	0.159 ^a	0.026	0.117 ^b	0.022	0.129 ^c	0.013	0.133 ^c	0.018	0.101 ^d	0.015
21.06.2005	0.132 ^a	0.020	0.099 ^b	0.019	0.109 ^c	0.014	0.111 ^c	0.016	0.079 ^d	0.014
22.06.2005	0.110 ^a	0.020	0.079 ^b	0.018	0.082 ^{bc}	0.011	0.088 ^c	0.015	0.062 ^d	0.013

Superscript letters (a, b, c) indicate significant ($p<0.05$) differences of the means between sites for each sampling date.

*Due to rounding of the last digit, significant difference exists between ungrazed 1999 and winter grazing, but not between continuous grazing and winter grazing.

1.4.3 Spatial dynamics of soil moisture as subject to grazing intensity

The spatial characteristics of soil moisture distribution on each site were assessed by means of a geostatistical analysis. In the variogram function, the sill indicates the in-field variability of soil moisture. Hence, a high amplitude of soil moisture values results in a high sill value. The box-plots in Fig. 1.6 show the variance of sill values during the sampling period.

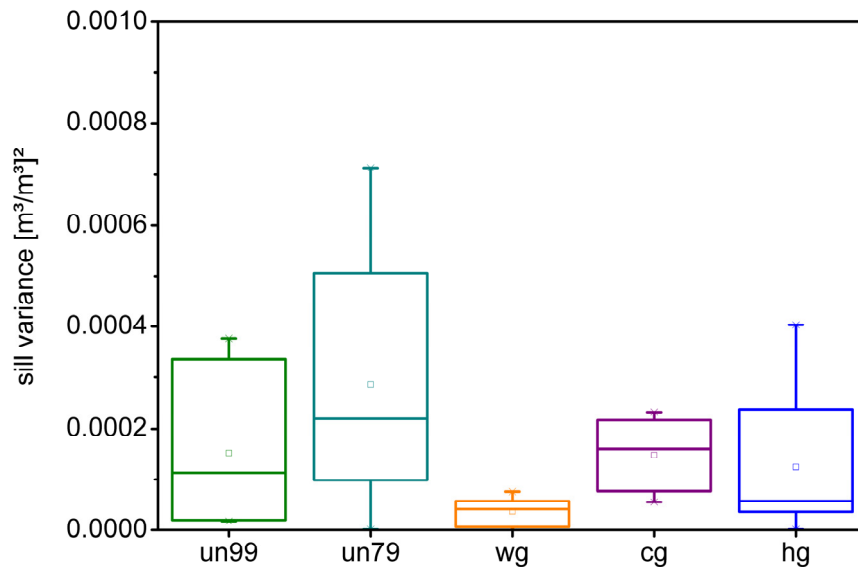


Figure 1.6. Box-plots showing the range of sill variance during the sampling period for each grazing treatment (Schneider et al. 2008, see chapter 3.3.1).

The small-scale variance as indicated by the nugget is excluded in this graph, so field-scale variability of soil moisture solely determines the shape of the box-plots. The ug79 site shows the highest range of sill values, whereas the wg site shows a low range of sill values, indicating high spatial variability of soil moisture on the first and low spatial variability on the latter site. Interpolated maps showing spatial distribution of soil moisture on the five sites for all sampling dates are given in chapter 3.3.1 (Fig. 3.6). As topographical or soil characteristics are very similar to the ug99 site (Steffens et al. 2008), the high range of sill values on the ug79 site has to be explained with its vegetation characteristics, which is a mixture of brush-like and grass vegetation. The shrubs impede infiltration of precipitation through interception and hence cause patterns of dryer and wetter patches. In consequence, spatial heterogeneity and the difference between low and high soil moisture values are highest on this site. In general, sill values decreased along with decreasing mean soil moisture (see chapter 3.3.1). Thus, soil moisture distribution is homogenised during dry periods and becomes more heterogeneous when wetness of the top soil increases.

1.4.4 Conceptual understanding of soil moisture dynamics in the study area

Although the soil moisture measurements on the experimental sites showed differences, the effects of grazing on soil moisture dynamics remain ambiguous. Absolute differences between the sites are moderate and neither temporal nor spatial characteristics show a clear fingerprint of grazing impact. Although the ug99 and hg sites are significantly different throughout the sampling period, the two ungrazed sites, too, show significant differences. In consequence, the observations do not allow for a classification of grazing effects on soil moisture; i.e., ungrazed sites do not necessarily hold best soil water storage properties. The results demand for a careful interpretation of grazing effects on environmental processes. As has been discussed in chapter 1.3, evapotranspiration is a major component of hydrological fluxes during the vegetation period. Therefore, vegetation characteristics and soil cover control the degree of evaporation, transpiration and interception and thus, the vertical fluxes of water. Based on vegetation studies by Fan et al. (2008), the conceptual model to explain the moderate differences in soil moisture observed is shown in Fig. 1.7.

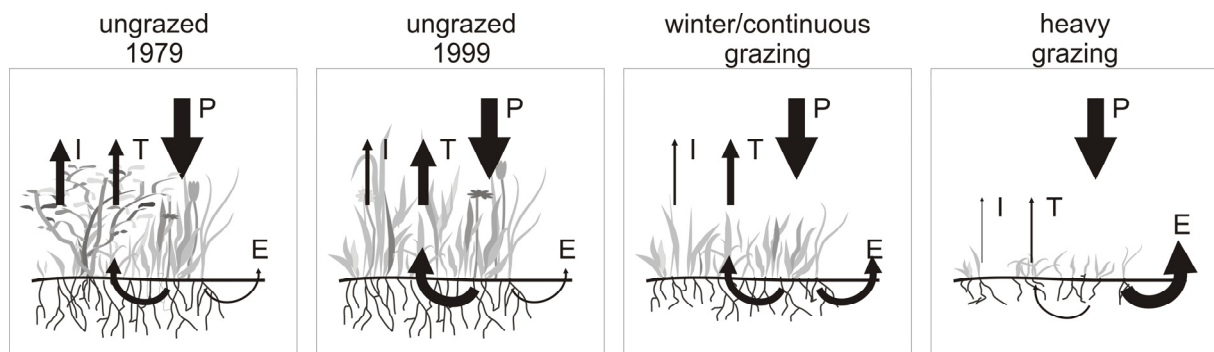


Figure 1.7. Conceptual model of plot scale vertical hydrological fluxes at natural and overgrazed sites of Inner Mongolian steppe ecosystems (P, precipitation; E, evaporation; I, interception; T, transpiration) (Schneider et al. 2008, see chapter 3.3.2).

Due to high LAI, biomass and litter cover on the ug79 and ug99 sites (Fan et al., 2008), interception and transpiration dominate, while evaporation from bare soil is low on these sites. The share of brush-like vegetation on the ug79 site might favour interception and causes less infiltration than on the ug99 site. The share of evaporation and transpiration is balanced on the wg and cg sites, as litter and vegetation cover is reduced. Hence, higher infiltration and lower interception rates lead to a higher share of soil water available for transpiration and evaporation. Low vegetation and litter cover on the hg site results in reduced interception and transpiration rates on the one hand, and raised evaporation rates on the other hand. The lessons learned from this study need to be considered when land use effects on soil water fluxes are implemented in hydrological models.

1.5 Reducing measurement efforts: time-stable points and their application in remote sensing validation

1.5.1 The time-stability concept

Despite the importance of soil moisture data in environmental studies, measuring it in a sufficient temporal and spatial resolution is labour intensive. Soil moisture studies need to be designed according to the number of data points that are required to reliably estimate mean (or minimum; maximum) soil moisture of a given area. Particularly for long-term studies, the question arises whether representative sampling points exist that reproduce mean soil moisture (or other statistical values) of a larger sampling volume with low variance during multiple samplings and to which extent spaceborne observations are useful in mesoscale soil moisture studies.

A technique to reduce the number of soil moisture sampling points was introduced by Vachaud et al. (1985) with the term *temporal stability*. The soil moisture measurements are analysed with respect to their statistical characteristics throughout multiple sampling dates. A sampling point is rated to be time-stable when it is close to mean field soil moisture throughout the sampling period, and when the deviation from mean field soil moisture does not vary a lot between the sampling dates. Mean deviation $\bar{\delta}_i$ of a sampling point i from mean field soil moisture is calculated as

$$\bar{\delta}_i = \frac{1}{m} \sum_{j=1}^m \frac{S_{i,j} - \bar{S}_j}{\bar{S}_j} \quad (2)$$

$S_{i,j}$ is soil moisture at sampling point i and sampling date j , \bar{S}_j is mean field soil moisture at sampling date j , and m is the number of sampling dates. Close proximity of a sampling point at location i to mean field soil moisture is indicated if $\bar{\delta}_i$ is close to 0. Temporal stability of a sampling point is best when the standard deviation of $\bar{\delta}_i$ is low.

The method was tested with the soil moisture data raised in the course of the three-year study. Detailed results are presented in chapter 4. Time-stable properties were calculated for all sampling points on the ug99, ug79, wg and hg sites. Identification and application of time-stable points allow monitoring soil moisture with only few sampling points continuously with a known precision. The error related to the prediction can be assessed by varying the number of time stable sampling points in order to find an optimum layout for the requirements of the experiment.

Apart from reducing the measurement effort, time-stable points may be useful to validate satellite soil moisture products. The study catchment covers an area of 3600 km² and is difficult to access in wide parts. Remoteness on the one hand, and data scarcity on the other hand demand for approaches to enhance our hydrological understanding without requiring intensive field work. Remote sensing is a promising tool in hydrology since various systems cover a wide range of information.

In this study, soil moisture data derived from ERS scatterometer products with a spatial resolution of 50 km were used. The backscatter signal of the active microwave sensor depends on soil moisture content. Backscatter is lowest under dry conditions and highest under wet conditions. ERS scatterometer data used in this study are distributed as relative surface wetness index (surfwet) for the top 0.05 m of

the soil. Using highest and lowest backscatter values, surfwet provides relative soil moisture, ranging from 0% (wilting point) to 100% (saturation) (Wagner et al., 1999). To obtain comparable data sets, minimum and maximum ground based measurements (i.e. after long drought and intensive precipitation) were used to convert the absolute soil moisture values into a wetness index ranging from 0% to 100% accordingly. Soil moisture measurements from 2004, 2005, and 2006 were applied to analyse time-stable characteristics and to compare ground and spaceborne soil moisture data.

1.5.2 Temporal stability of soil moisture on the grazing sites

Time-stable points were identified on all sites. Nevertheless, differences occurred related to the number of points meeting certain quality criteria (Table 1.3). The hg site had the lowest number of sampling points remaining within 5% difference from mean in both, 2004 and 2005. Despite the low number of time-stable points within 5% error, the hg site showed highest temporal persistence on the field scale as indicated by a rank correlation analysis (see chapter 4 for more details).

Table 1.3. Percentage of sample points located within 5% difference, and 1 and 2 standard deviations (std) from mean (Schneider et al., 2008, see chapter 4.3).

	ug79		ug99		wg		hg	
	2004	2005	2004	2005	2004	2005	2004	2005
5%	50	48	54	54	74	59	22	32
1std	95	97	94	92	97	93	81	70
2std	100	100	100	100	100	100	98	95

The best points were chosen depending on the proximity of $\bar{\delta}_i$ to 0, and low standard deviation. Soil moisture measurements of the four best time-stable points on each site calculated from the 2004 data were chosen to compare with mean field soil moisture in 2005 (Fig. 1.8). On all sites, the time-stable points match mean field soil moisture within two times the standard deviation and reproduce soil moisture dynamics throughout the sampling period in 2005. The mean value of the four best time-stable points is very close to mean field soil moisture calculated from all sampling points. This implies that using the average of only few time-stable points yields very precise estimates of mean field soil moisture.

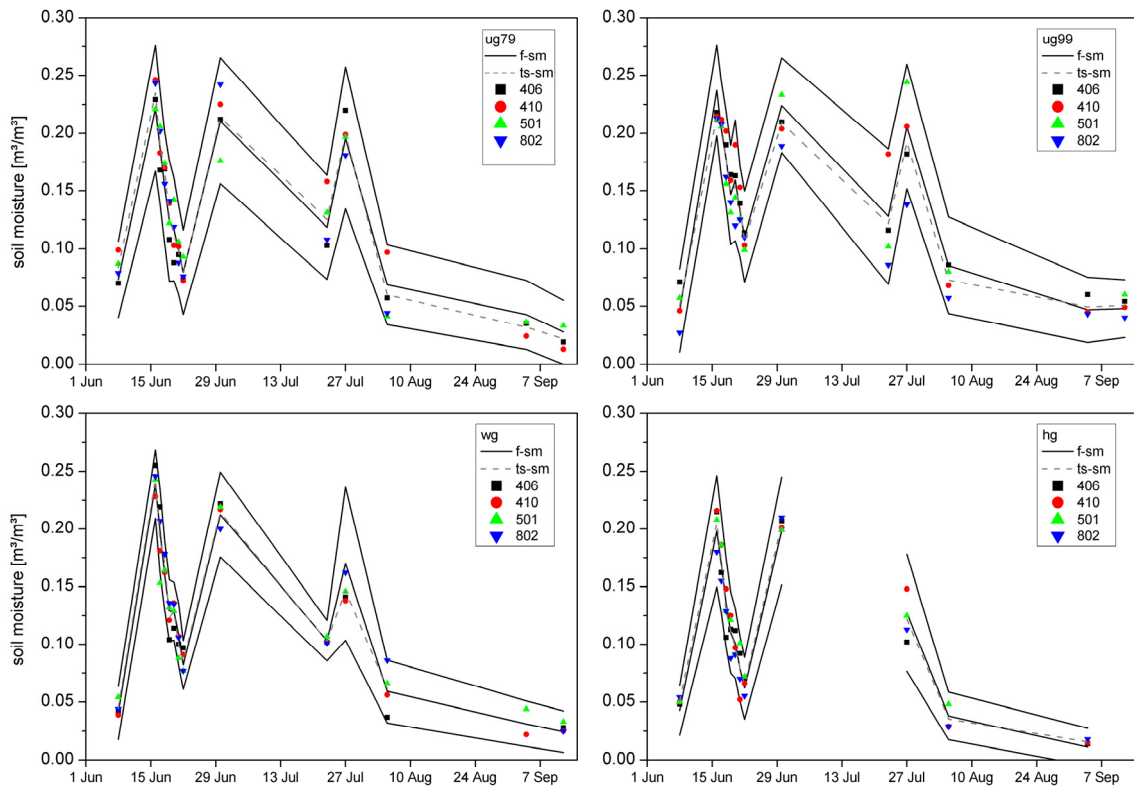


Figure 1.8. Soil moisture dynamics on the four grazing sites in 2005. Dashed line: mean soil moisture calculated from time stable points, solid line: field mean soil moisture and $\pm 2\sigma$ (Schneider et al. 2008, see chapter 4.3).

In multi-year studies it is important to know whether time-stable points retain their characteristics beyond a single vegetation period. Time-stable points were calculated from 2004 data, from 2005 data, and from a combination of both years. The quality of the selected time-stable points was tested with the 2006 soil moisture data (see chapter 4.3, Fig 4.4). In some cases, time-stable points changed from 2004 to 2005 which is mainly related to higher standard deviations, while other points kept time-stable characteristics over both years. The root mean squared error (RMSE) between mean soil moisture of the four best time-stable points and mean soil moisture of all sampling points of the 2006 data is given in Table 1.4.

Table 1.4. Mean RMSE between soil moisture of time-stable points and the 2006 mean field soil moisture (Schneider et al., 2008, see chapter 4.3).

	Best time-stable points selected from 2004 data	Best time-stable points selected from 2005 data	Best time-stable points selected from 2004+2005 data	Worst time-stable points selected from 2004 data
ug79	0.008	0.010	0.009	0.015
ug99	0.006	0.010	0.009	0.015
wg	0.006	0.006	0.004	0.024
hg	0.007	0.006	0.006	0.019

Although RMSE is low in both, 2004 and 2005, the values slightly increase in 2005. The precipitation conditions in 2004 and 2006 were similar, while 2005 was much dryer than average. The different climate conditions seem to affect the applicability of the 2005 time-stable points in 2006. The combination of the 2004 and 2005 data did not improve the quality of the time-stable points, indicating that one vegetation period is sufficient for deriving time stable characteristics.

In order to optimise the sampling strategy without serious loss of information, a balance between the number of sampling points and the accuracy of the predictions is required. Using only one time-stable point might promote errors related to the reproducibility of the measurement. As shown in Fig. 1.9, averaging several time-stable points lowers RMSE considerably. On some sites, RMSE is reduced by almost 50% if three or four points instead of only one point are applied.

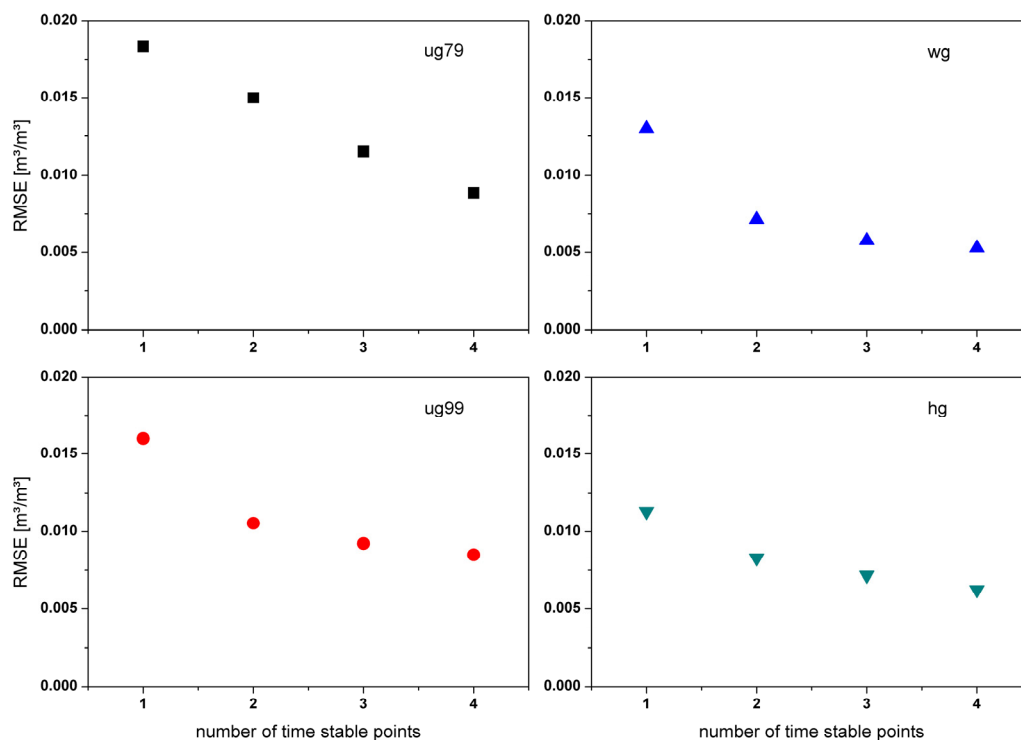


Figure 1.9. Effect of number of time-stable points on RMSE values (Schneider et al. 2008, see chapter 4.3).

1.5.3 Application of time-stable points in remote sensing

The ground based soil moisture measurements were converted into a relative soil moisture index and compared to ERS scatterometer surface wetness index. As the four grazing sites represent the mixture of grazing management present in the study area, the field measurements were aggregated. Even then, ground-based measurements have a much smaller footprint than the ERS footprint. Mean soil moisture and standard deviation of all sampling points and of the time-stable points is

compared with the ERS soil moisture data in Fig. 1.10. The ERS data are split into three categories: *ERS closest* is the pixel with its center closest to the experimental sites, *ERS amean* is the areal mean and *ERS wmean* is the weighted mean (i.e. inverse distance weighted of the pixel center) of the four pixels surrounding the experimental sites.

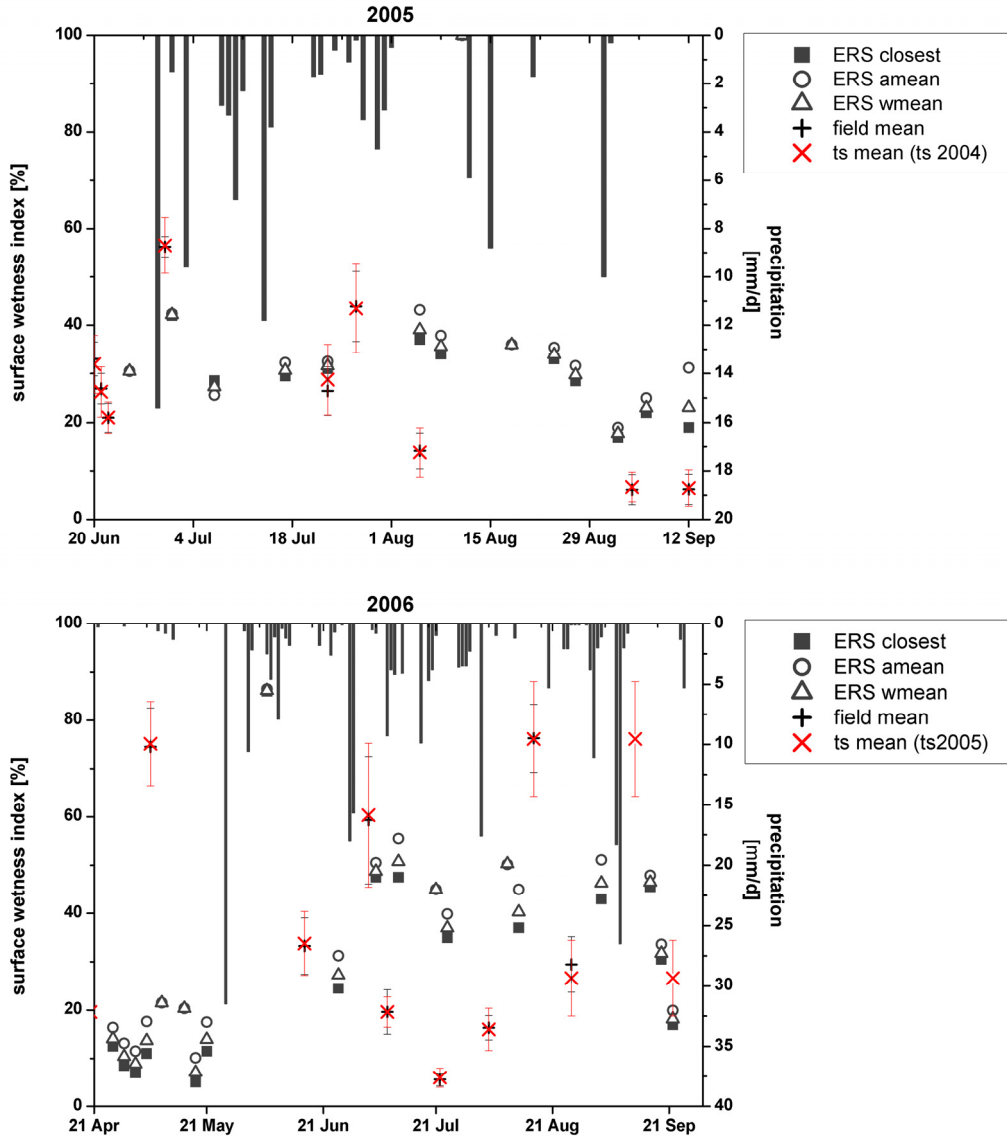


Figure 1.10. Comparison of ERS soil moisture and ground-based measurements in 2005 and 2006. ERS closest: single value of pixel closest to grazing sites; ERS amean: mean values of four pixels adjacent to grazing sites; ERS wmean: inverse distance weighted mean values of four pixels adjacent to grazing sites; field mean: mean soil moisture of all sampling points on the four grazing sites; ts mean: mean soil moisture of selected time-stable points on the four grazing sites. Error bars indicate one standard deviation (Schneider et al. 2008, see chapter 4.3).

Time-stable point soil moisture is within the range of the aggregated field mean soil moisture. Hence, the information gained from time-stable points allow upscaling to field scale. On the other hand, ground-based field soil moisture and ERS data at a much larger scale match in some cases, but differ considerably in other cases. The grazing sites represent the range of grazing management in the Xilin river catchment, and no abrupt land use changes occur which may introduce changes in roughness length. Therefore, precipitation patterns may account for the differences. During the vegetation period, precipitation is mainly of convective type. This may lead to local patterns which can not be captured within the 50 km footprint of the ERS sensor and hence introduce huge bias between ground-based and spaceborn measurements.

1.5.4 Conclusions

Time-stable sampling locations were identified on all grazing sites and indicated the potential to reduce measurements efforts on field scale experiments considerably. Data from one vegetation period suffices to derive time-stable points for long-term studies. Additional measurements in the following year did not improve the performance of time-stable points. Yet, precipitation conditions differed between these two years and the question remains whether calculating time-stable points from longer time series would improve their performance. The number of time-stable points necessary for acceptable predictions of mean field soil moisture needs to be chosen carefully, as the error increases considerably if only one or two time-stable sampling locations are applied.

The validation of remote sensing soil moisture data with time-stable sampling locations proved to be difficult even if field scale soil moisture was captured satisfactorily. The bias is attributed to the large scale gap between the field scale measurements and the footprint of the satellite sensors. Convective precipitation cells cause local patterns of soil moisture which can only be captured by sensors with a higher spatial resolution. Although the ERS scatterometer soil moisture data is only partly beneficial to complement hydrological information in the data sparse Xilin river catchment, the next generation of sensors with a higher spatial resolution might reduce the current restrictions.

1.6 Synthesis and outlook

The aim of this study was to assess the impact of grazing intensity on hydrological fluxes in a data sparse, remote catchment in a steppe environment in northern China's Autonomous Region Inner Mongolia. Soil moisture data were assembled at various spatial scales with ground-based and space born methods. The approach provided detailed information on grazing effects on surface soil moisture on the one hand, and on moisture dynamics on the catchment scale on the other hand. Both aspects increased our understanding of dominant hydrological processes in the catchment, in particular on the exchange between soil, vegetation and atmosphere. It will further help when relevant processes need to be implemented in environmental models. This is an important step to allow for transferability to other regions where comparable environmental conditions drive hydrological processes.

Returning to the initial questions raised in chapter 1.1, some surprising results could be revealed. Although grazing does change the water storage in the top soil to some degree, the differences remain moderate. Even more important, two ungrazed sites reveal differences that were in the same range as the differences found between an ungrazed (ug99) and a heavily grazed site. This implies that singular information “ungrazed vs. grazed” may lead to misinterpretations or biased results when translated into hydrological models, or is used for decision making.

If we try to transfer the study results on soil moisture dynamics on a larger model scale, we need to consider the influence of atmospheric forcing on hydrological fluxes. On the field scale, climate conditions are steady, and hence, the grazing, or more general, the land use footprint will become obvious. On larger scales, however, atmospheric forcing superimposes small scale land use effects. In addition, with increasing scale, changes in soil, vegetation and topography are likely to become more pronounced. Therefore, the effect of grazing might be inferior compared to the overall variability of factors influencing hydrological processes. The study scale will determine to which degree generalization is acceptable in a model. Nested approaches with detailed input data at small scales and lumped information at larger scales are a key to balance feasibility with the demand for sufficiently detailed input information.

Remote sensing is a valuable tool to assess information for hydrological applications on large scales. Yet, there are restrictions to which degree different grazing intensities and related vegetation characteristics can be resolved by a sensor. For example, spectral differences of various grazing intensities might be too small to allow for reliable grazing classification with optical sensors. Although soil moisture data derived from the ERS scatterometer provide information on meso- to macroscale, the information is restricted as only the top 0.05 m are penetrated by the radar. Information for soil profiles can only be derived with transfer functions, which require calibration on the environmental conditions in the catchment. Furthermore, like in the Xilin catchment, the scatterometer footprint might be larger than the precipitation footprint in case convective precipitation is dominant. Therefore, soil moisture dynamics and spatial patterns that might lead to spatially heterogeneous runoff generation can not be captured by the sensor. The results show that neither too detailed information on the one hand, nor data on a high level of generalisation can solely provide sufficient information to understand hydrological processes in a remote, data sparse catchment. Moreover, combined approaches are required.

Although surface soil moisture measurements reflect precipitation events, their contribution to runoff generation is small. During the vegetation period, hydrology in the Xilin river is strongly influenced by evapotranspiration, and precipitation returns to the atmosphere quickly. In consequence, runoff generation via surface runoff, subsurface flow and percolation is limited to events where precipitation exceeds the rate of evapotranspiration. Such events are rare and occurred only two to three times within the three years study period, indicating that other sources feed the Xilin river. Groundwater that is formed during snowmelt (or is entering the catchment from outside its surface boundaries) might contribute substantially to runoff generation. Hydrological simulations failed to predict hydrological flows satisfactorily. Therefore, approaches beyond mere rainfall-runoff modelling are required. These include tracer studies on a subcatchment to catchment scale to clarify the role of snowmelt, precipitation and groundwater contribution to runoff generation. In addition, transit time of water through the catchment can be studied with isotope measurements. In

combination with standard hydrological modelling approaches and additional information from remote sensing (e.g. land use or vegetation characteristics), a multi-scale and multi-tool approach is recommended to improve our conceptual understanding of the (with respect to hydrology) poorly studied Xilin catchment. In the second phase of the MAGIM project, open questions concerning runoff generation will be studied with such an approach.

One may raise the questions whether hydrological studies in a remote environment in northern China matter at all. At first glance this may be true – at second glance, however, we have to consider the global role of steppe ecosystems with respect to the ecosystem functions they provide, and environmental and social sustainability. Grassland and dryland environments cover huge areas worldwide and constitute an important reservoir for carbon sequestration (Ojima et al., 1993). Availability of water controls the release or uptake of carbon, e.g. as gaseous compounds, from or into these reservoirs. Climate change or land use management affect environmental conditions and, in consequence, alter the exchange rates between soil, vegetation and atmosphere. Investigating the role of water fluxes therefore is essential to estimate matter budgets in steppe ecosystems.

Being an area with a long settlement history, the Xilin river catchment is the social and economic basis for the local population. Settlement politics and increasing wealth raise the pressure on ecosystem services in the catchment. For example, growing herds require more space which causes property conflicts; agricultural branches such as milk farming have been introduced to the area recently, although natural conditions in the Xilin river catchment do not favour intensive forms of animal breeding. Population growth in the area will further increase the demand for water resources. The situation becomes even more complicated when we consider climate variability in the region affecting biomass and fodder production. Both, amount and onset of the rainy season is highly variable in the Xilin river catchment. This is an important factor when grazing management is considered: under the given climate conditions, inflexible grazing management either results in reduced efficiency (i.e. productivity), or promotes environmental degradation. The study showed that even if the total amount of evapotranspiration is not affected by grazing, there is a considerable effect on plant productivity. As the share of evaporation, transpiration and interception is influenced by grazing intensity (Fig. 1.7), a management system that promotes the transpiration component (and hence, biomass production) will achieve sustainability between plant and livestock production. Neither completely ungrazed systems nor heavily (over-)grazed systems can provide the required functions under both, environmental and socioeconomic aspects. The implications of natural factors and land use on socio-economic development in the Xilin river go beyond the scope of the presented work. Nevertheless, the study provides one piece in the puzzle of interacting environmental factors of a fragile ecosystem.

2 Evaluation of evapotranspiration methods for model validation in a semi-arid watershed in northern China

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Abstract

This study evaluates the performance of four evapotranspiration methods (Priestley-Taylor, Penman-Monteith, Hargreaves and Makkink) of differing complexity in a semi-arid environment in north China. The results are compared to observed water vapour fluxes derived from eddy flux measurements. The analysis became necessary after discharge simulations using an automatically calibrated version of the Soil and Water Assessment Tool (SWAT) failed to reproduce runoff measurements. Although the study area receives most of the annual rainfall during the vegetation period, high temperatures can cause water scarcity. We investigate which evapotranspiration method is most suitable for this environment and whether the model performance of SWAT can be improved with the most adequate evapotranspiration method.

The evapotranspiration models were tested in two consecutive years with different rainfall amounts. In general, the simple Hargreaves and Makkink equations outmatch the more complex Priestley-Taylor and Penman-Monteith methods, although their performance depended on water availability. Effects on the quality of SWAT runoff simulations, however, remained minor. Although evapotranspiration is an important process in the hydrology of this steppe environment, our analysis indicates that other driving factors still need to be identified to improve SWAT simulations.

2.1 Introduction

Measuring and modelling key features of the hydrology of semi-arid watersheds can hold unexpected challenges as compared to similar work developed in humid environments. Precipitation and temperature patterns differ, which when combined with additional differences in soil and vegetation properties, lead to a significant shift in the distribution of runoff processes. In humid regions, relatively low intensity long duration precipitation over soils with relatively high infiltration capacities lead to a characteristic pattern of downslope wetting and a high likelihood of lateral subsurface flows. While there is considerable evidence to suggest that the degree of connectivity and convergence is not as high as one might predict based upon the topographic index (Western et al., 1999; Güntner et al., 2004), various simulations relying on the concept of connected lateral subsurface flows have been successfully developed in wet catchments. These include a wide variety of applications using the TOPMODEL

concept (where the classic example is Beven and Kirkby (1979)) or Dupuit Forchheimer theory (Wigmosta et al., 1994; Vaché and McDonell, 2006). In addition, simulations using the semi-distributed SWAT (Soil and Water Assessment Tool) model were shown to improve, for the Dill catchment, Germany, with the explicit incorporation of lateral subsurface flows (Eckhardt et al., 2002). In more arid regions, the degree of connectivity declines, process non-linearity increases and models that perform acceptably in humid environments may need to be adapted. A key component of this adaptation is the realization that evapotranspiration (ET) may play a significantly larger role in the water balance of semi-arid catchments, and may therefore be a focus of catchment simulations in these regions.

In our study catchment, the Xilin River, in Inner Mongolia, PR China, runoff ratios are extremely low and ET flux measurements are relatively high, suggesting that finding an appropriate ET method is a key component in the development of any hydrological model.

Many equations to model ET are available. One of the most common used ones is the Penman-Monteith (PM) formula. It is recommended by the FAO (Allen et al., 1998) as reference ET (adapted from grass ET when water is not limited). Nevertheless ET often has to be estimated under water stress conditions. Several studies investigated which method is most suitable for semi-arid areas (Frank, 2003; DehghaniSanij et al., 2004; Kurc and Small, 2004). Liu and Erda (2005) compared the Priestley-Taylor (PT) method with the reference crop PM equation for six weather stations in the semi-arid northern China. As PT only delivers acceptable results under certain conditions, they recommend the PM equation. This is in agreement with López-Urrea et al. (2006), who tested seven ET methods in semi-arid regions in Spain. They found the PM method to be most suitable, but underline that also simpler methods (e.g. Hargreaves) performed surprisingly well.

Preliminary model results for the present hydrological case study of the Xilin river catchment using the Soil Water Assessment Tool (SWAT, (Arnold et al., 1998)) are unsatisfying. The model structure is unable to capture (1) the spring snowmelt peak and (2) summer discharge. Modelled spring snowmelt is far too low as compared to observed data, and summer peaks are considerably higher than observed under extreme precipitation events, but much lower under average conditions. Temporal dynamics, as well as annual discharge, are not captured. As neither manual nor automatic calibration improved the model performance to an acceptable degree we seek alternative ways to understand and finally simulate the hydrological processes in this semi-arid catchment. In a first step we focus on the summer discharge. Due to high temperatures, the watershed can be water limited even during the wet vegetation period. Thus evaporation and transpiration magnitudes are a key component of the hydrologic cycle. To evaluate the potential role of ET simulation errors in the SWAT results, we independently analyse the performance of several ET methods, all of which are supplied within the model. The modelled ET rates are compared to water fluxes measured by the eddy covariance technique.

2.2 Materials and methods

2.2.1 Study area

The study site is located approximately 400 km north of Beijing at the southern rim of the Xilin river catchment (3650 km²) in the Province of Inner Mongolia (PR China) (Fig. 2.1). It belongs to the Eurasian steppe ecosystem and is marked by a continental climate (Fig. 2.2b). The mean monthly temperature amplitude ranges between 18°C (July) to -23°C (January). Mean annual precipitation is 350 mm, but is highly variable between 150 to 500 mm. The wet season from June to August receives 60-80% of annual rainfall (Chen, 1988); nevertheless precipitation during the vegetation period differs among years (Xiao et al., 1995).

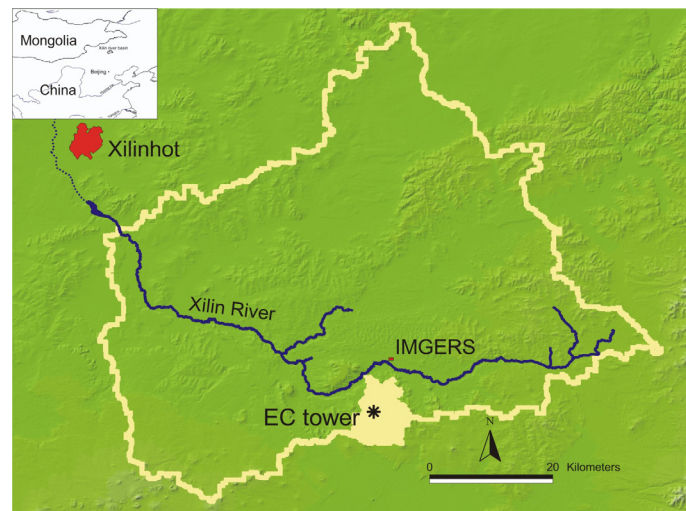


Figure 2.1. Outline of the Xilin river catchment with the location of the eddy flux station (EC tower) situated in the subcatchment relevant for the study. IMGERS: Inner Mongolia Grassland Ecosystem Research Station.

In contrast to the distinct precipitation peak during the summer months, the hydrograph does not reflect the annual precipitation cycle: despite low snow rates during the winter months (November to March) vernal discharge reaches highest values during the melt period, whereas the precipitation peak in June and July does not result in high discharge (Fig. 2.2a).

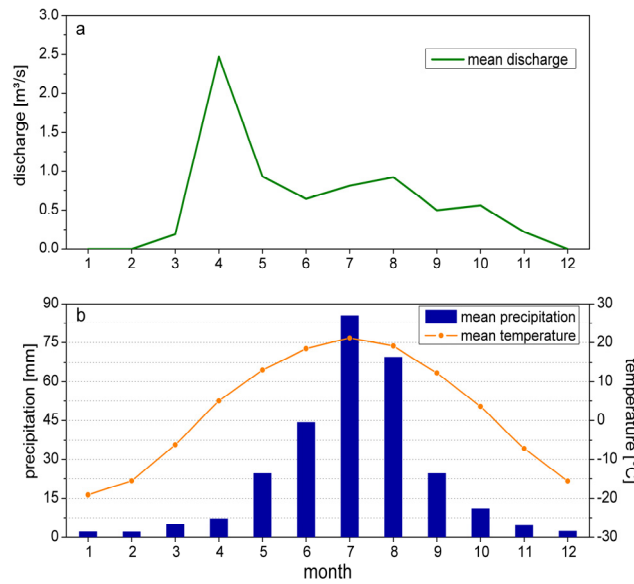


Figure 2.2. Mean discharge (a) and precipitation and temperature (b) in the Xilin river catchment.

2.2.2 Model and observational data

The SWAT model

SWAT is a semi-distributed ecohydrological model. Water fluxes are calculated for unique soil and land use combinations (Hydrological Response Units, HRU) within a subcatchment. After deriving surface runoff and infiltration processes, evaporation and transpiration are calculated, thus reducing the available water for percolation or river runoff. The fluxes calculated for each HRU are accumulated within the relevant subcatchment and then routed along the stream network to the watershed outlet.

Several methods with varying data requirements for evapotranspiration estimation are incorporated in SWAT: the rather complex Penman-Monteith (PM) and Priestley-Taylor (PT) methods, and the simpler Hargreaves (HG) formula (see Neitsch et al. (2001) for details). The modified SWAT-G version, which is used for all model runs in this study (Eckhardt et al., 2002), was further extended by the Makkink (MK) method, a simplified version of the PT formula (Makkink, 1957; De Bruin and Lablans, 1998). The parameters required for the ET models are given in Table 2.1.

Table 2.1. Data requirements of the four ET models.

Parameter	PM	PT	HG	MK
Air temperature	X	X	X	X
Relative humidity	X	X		X
Solar radiation	X	X		X
Wind speed	X			

In SWAT, the ET methods estimate potential ET (PET) as a first step. Actual ET (AET) is then derived from PET as a function of plant parameters and water storage in the soil (Neitsch et al., 2001). In the present work all ET methods were applied in an uncalibrated mode.

Eddy flux measurements

Evapotranspiration rates were derived from measurements of an eddy flux station located on experimental sites in the southern part of the catchment. The change of water vapour concentration was measured with an open path infrared gas analyser (LI-7500, LI-COR Inc., Lincoln, Nebraska, USA). In combination with wind speed and wind direction measurements (CSAT3, Campbell Scientific Inc., UK), an eddy covariance approach was applied to calculate latent heat fluxes. Actual evapotranspiration was derived from this calculation.

The observed and modelled evapotranspiration rates are compared for the summer of 2004 (16th August – 30th September) and 2005 (15th May – 24th September). Though SWAT was run for the 3600 km² catchment, we only compared measured AET with modelled AET data from the subcatchment where the eddy flux tower is located (Fig. 2.1). Precipitation during the study period was measured at the eddy flux station. This data is assigned to the subcatchment in the SWAT model, so the comparison is based on uniform precipitation data. Precipitation for the rainy season from May to September was 288 mm in 2004 and 125 mm in 2005 (measured at Inner Mongolia Grassland Research Station, see Fig. 2.1). The comparison of the different ET methods thus also comprises an evaluation on how the methods perform under varying boundary conditions (normal vs. dry year).

2.3 Results and discussion

2.3.1 Comparison of summer sums

Observed and calculated ET sums for 2004 and 2005 are given in Table 2.2. Summer precipitation was much higher in 2004 than in 2005. In 2004, PT and PM underestimate, and HG and MK overestimate measured ET.

Table 2.2. Summary of precipitation and observed and calculated ET for 2004 and 2005.

	Rainfall	Observed ET	PT	PM	HG	MK
16.8. – 30.9.2004	82.6	99.3	92.1	81.3	107.4	104.3
Deviation from observed (%)			-7	-18	8	5
15.5. – 24.9.2005	113.1	174.5	140.0	145.0	164.0	155.3
Deviation from observed (%)			-19.8	-16.9	-6	-11.2

Calculated ET does not fit observed ET exactly, yet the deviation from measurements is moderate. Though the MK method shows lowest deviation from measured ET, all methods perform in an acceptable way. In contrast, we found higher differences in the much dryer summer of 2005. During a very dry period from 16th August to 24th September with only 5 mm of precipitation observed and calculated ET rates show marked deviations which reach up to 40% of the observed ET values (data not shown).

The deviations from observed ET are lower when the complete measurement period 2005 (15th May – 24th September) is considered. The models seem to smoothen out extremes throughout the vegetation period. Yet, all ET methods underestimate observed ET when the complete monitoring period of 2005 is analysed, but yield overestimations during the very dry period. Overestimation occurs more pronounced under dry conditions and deviations from observed data tend to rise under high water deficits.

2.3.2 Comparison of daily ET

Fig 2.3a-d compares observed and modelled ET rates in daily time steps. In 2004 all models reflect observations in an acceptable way, though PM partly underestimates and HG and MK partly overestimate measurements.

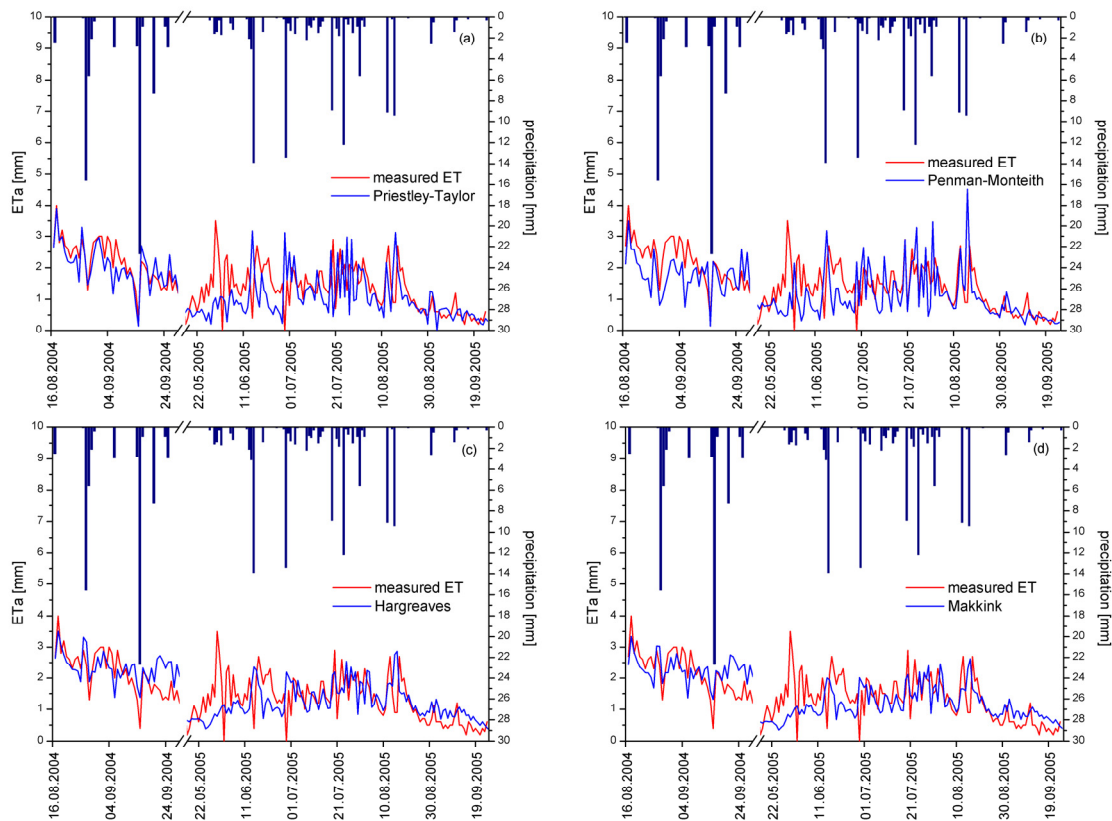


Figure 2.3. Observed vs. measured actual evapotranspiration (ETa): (a) Priestley-Taylor, (b) Penman-Monteith, (c) Hargreaves, (d) Makkink.

Both, observed and modelled ET rates are on a lower level in 2005 than in the previous year. Except for the PM model, none of the ET methods show high anomalies from observed ET values. The more complex PT and PM models show higher amplitudes as compared to observational data or the simpler HG and the MK methods. None of the models produces systematic under- or overestimation compared to observed data, and fluctuate above or beneath the observed data.

The ability of the four ET methods to reproduce daily ET measurements was assessed with the Nash-Sutcliffe-Efficiency (NSE) (Nash and Sutcliffe, 1970). The closer NSE approaches 1, the better modelled and observed values match. Table 2.3 shows the results for 2004 and 2005. Though the MK method has lowest deviation from observed in 2004, only the PT method reaches a NSE of more than 0.5. The quality of the calculations clearly decreases in 2005, as all methods have negative NSE values. PT and PM perform particularly bad in 2005. Our results suggest that the simpler HG and MK methods are superior to the more complex PT and PM methods. The MK method performs unexpectedly well considering that it was developed under temperate humid conditions.

Table 2.3. Quality of ET simulations in 2004 and 2005 (NSE: Nash-Sutcliffe-Efficiency, RMSE: Root Mean Squared Error).

	PT		PM		HG		MK	
	2004	2005	2004	2005	2004	2005	2004	2005
NSE	0.648	-0.636	-0.001	-0.474	0.285	-0.028	0.275	-0.004
RMSE	0.413	0.879	0.696	0.834	0.588	0.697	0.593	0.688

2.3.3 Influence of ET method on SWAT output

Though the four ET methods yield different results, none of them leads to satisfying predictions of discharge with the SWAT model. Table 2.4 summarises observed and simulated mean discharge sums for the vegetation period. All simulations underestimate discharge and fail to reproduce observations. Even the “best” simulation using the MK method estimates only one third of actual discharge. None of the ET methods clearly improves the SWAT simulations. In consequence, other processes might influence model performance to a greater extent than the chosen ET method does.

Table 2.4. Mean accumulated summer discharge (1st May – 30th Sep) calculated from observed data and SWAT simulations with four ET methods. Values are calculated from 8 consecutive years with available data.

	Summer discharge [m ³ /s]	Fraction of observed discharge
Observed	87.12	-
Simulated with Priestley-Taylor	12.54	0.14
Simulated with Penman-Monteith	14.88	0.17
Simulated with Hargreaves	24.82	0.28
Simulated with Makkink	27.65	0.32

2.4 Conclusions

In semi-arid regions, ET is a large component of the hydrologic cycle, and a key component of any applied catchment model. In an effort to quantify the potential effect of ET estimation on a model focused on runoff generation, we evaluated four ET methods and compared their performance with observational data. Considering the uncertainties associated with modelling and measuring ET, we conclude: (1) The different methods did not reflect observed sums for 2004 and 2005 accurately; nevertheless, deviations remained moderate. (2) The quality of the ET simulations varied depending on water availability. Especially during very dry periods, the HG and MK methods showed a high bias, whereas they performed better than the PT and PM methods when periods with mixed wet and dry conditions are considered. (3) The ability to capture day-to-day characteristics of measured data is also dependent on water availability. In general, the simpler HG and MK equations outperform the PM method in the study environment. While ET results differed between methods, effects on the simulated discharge response appears to be minor. We conclude from this observation that while ET is a key component of water cycle in this region, other factors also contribute to the inability of SWAT to capture the measured discharge response. These factors remain to be determined, but may include uncertainties in the spatial distribution of convective precipitation or deep groundwater recharge.

Acknowledgements

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3 Ambiguous effects of grazing intensity on surface soil moisture – a geostatistical case study from a steppe environment in Inner Mongolia, PR China

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Abstract

Does grazing intensity lead to changes in the mechanisms of water distribution and water storage in the topsoil? We attempt to answer this question by comparing temporal and spatial soil moisture variation on sites with different grazing intensities ranging from ungrazed to heavily grazed. The research was carried out in a semi-arid steppe environment in northern China. During the growing season, evapotranspiration rates usually exceed the amount of rainfall in these areas, so the ability to store water in the plant-soil system is crucial for biomass growth. In June 2005, the spatial variation of soil moisture in five grazing treatments (two ungrazed sites, one winter grazing, one continuous grazing and one heavy grazing site) was monitored on seven consecutive days after a precipitation event. We used capacitance sensors to measure volumetric soil moisture down to a depth of 0.06 m. The measurements were analysed with respect to the spatial correlation of data points (variogram analysis and kriging interpolation). The five sites did reveal some spatial patterns, but these were not very distinct. The capability of retaining water in the topsoil declined along the grazing gradient from ungrazed (wettest site) to heavily grazed (driest site). However, this pattern is disturbed, as the two ungrazed sites (ungrazed since 1979 and ungrazed since 1999) reacted differently to wetting and drying and showed significant differences on all sampling days.

Keywords: soil moisture dynamics, topsoil, geostatistics, grazing intensity, semi-arid grazing system

3.1 Introduction

China is one of the countries facing the most serious desertification problems in the world. A preliminary assessment of desertification in 1997 showed that the total area of affected land was approximately 2.6 million km², covering 27.3% of the total territory of China (Fukuo et al., 2001). Clearly, priority should be given to combat desertification processes in vast areas of China, including the province of Inner Mongolia, where grassland degradation is a key environmental problem.

Many factors contribute, interact and accelerate the degradation process of semi-arid grasslands. For example, the impact of management strategies on bearing capacity and grassland degradation is not adequately understood at present. This is due to the fact that depending on climatic conditions, soil type and vegetation structure contradicting results have been obtained. In some cases, grazing intensity had an impact on ecosystem performance whilst in other cases no effect was observed (Oba et al., 2000; Fuhlendorf et al., 2001; Vetter, 2005; Golodets and Boeken, 2006). Investigating the water-related interactions between plants, soil and atmosphere is a crucial step to better understand degradation processes that occur in semi-arid landscapes. Yet, only few investigations have dealt with the impact of grazing intensity on these interactions.

The distribution of soil moisture is controlled by several factors acting on various scales (Blöschl and Sivapalan, 1995). Controls on the plot scale, such as soil texture, bulk density and infiltration capacity, are superimposed by those on the field scale (e.g. vegetation pattern, topography) and regional scale (e.g. weather patterns). When local controls have been identified, they can be used to estimate soil moisture distribution. However, the predictive power of these factors is limited by the preliminary moisture condition of the research area. For example, Grayson et al. (1997) and Western et al. (1999) reported that spatial organisation of soil moisture varies largely between dry and wet states and that the organisation of patterns can have different causes under changing moisture conditions. Their soil moisture records showed a higher degree of spatial organisation under wet conditions than during dry periods.

Besides topography, vegetation patterns can also have a strong influence on the spatial pattern of soil moisture through root water uptake, interception and shading. In a comparison of burnt (i.e. lack of vegetation cover) and unburnt (i.e. with vegetation cover) catchments in a semi-arid environment in south-east Spain, Gómez-Plaza et al. (2001) showed that spatial distribution of soil moisture is related to the abundance of vegetation. When vegetation is scarce, soil moisture patterns are governed by local controls, such as texture. The soil moisture patterns could to a great deal be explained by these local factors, and, in accordance with the findings of Western et al. (1999), the correlation between the governing factors and soil moisture patterns was in general highest in wet conditions. In a different study, Archer et al. (2002) and Cantón et al. (2004) investigated the influence of vegetation characteristics on storage and vertical as well as lateral fluxes of water. In these studies, soil moisture regimes differed between open (inter-shrub) and shrub areas. Bushes were able to intercept rainfall with their aboveground biomass and guided water into the soil via their rooting system, which created pathways of preferential vertical flow. Depending on the depth of their rooting system, the bushes depended to a greater deal on rainfall (i.e. with a shallow rooting system) or on water coming from deeper soil layers (i.e. with deep rooting systems).

Besides the natural heterogeneity of soil properties and vegetation patterns, changes in these properties and patterns are often induced by human action (Fuhlendorf et al. 2001; Vetter 2005; Metzger et al., 2006; Navarro et al., 2006). Changes in land use and land management in semi-arid and arid grasslands can alter plant composition and finally plant cover and thus influence the storage or consumption of water in the soil. This is of particular concern in the Eurasian steppe ecosystems, where large-scale degradation of grassland has been reported (Christensen et al., 2004; Tong et al., 2004; Wang et al., 2006; Xu, 2006). Where the

vegetation cover is reduced by increased grazing activity, the soil is prone to wind and water erosion (Gao et al., 2002; Gao et al., 2003). In addition, grazing can substantially impact microclimatic processes for steppe environments in eastern Inner Mongolia, as has been shown by Li et al. (2000). They reported that factors such as albedo and roughness length are altered substantially on degraded or overgrazed areas, thus influencing the water budget in the topsoil. Chen et al. (2007) modelled the contribution of evaporation and transpiration to the overall amount of evapotranspiration under varying grazing impacts. They concluded that grazing favours evaporation, while transpiration is more dominant under non-grazing conditions. Yet, the overall evapotranspiration budget did not differ significantly under either grazing or non-grazing conditions.

To better understand the factors governing hydrological processes, information about spatial and temporal evolution of soil moisture is needed. Deriving conclusions from soil moisture measurements at only a few locations may result in large uncertainties, as soil moisture can be highly variable. A promising approach to reduce and quantify these uncertainties is the geostatistical sampling technique. Many authors have investigated the spatial and temporal variation of soil moisture (Famiglietti et al., 1998; Mohanty et al., 2000; Anctil et al., 2002; Huisman et al., 2002, amongst others). Depending on the research interest and thus on the measurement technique applied, the extent of the research area ranged from meters to kilometres, with either very detailed information on small-scale variability or general trends over large areas.

As has been discussed, soil water content is an important factor for ecological processes in the soil, vegetation and atmosphere. Due to changing boundary conditions as for example induced by management practices, soil water storage may be altered. In the present work, we therefore investigate the question of grazing impact on soil moisture evolution by the use of a geostatistical approach. The study has been conducted in a semi-arid steppe ecosystem of Inner Mongolia, PR China, where pronounced grassland degradation has been reported in recent years. The experimental set up allows a spatial analysis of soil moisture distribution and is the basis for a sound comparison of the impact of different grazing intensities on soil moisture. We hypothesize that (1) soil moisture dynamics depend on grazing intensity and (2) the ability to store water in the topsoil decreases with increasing grazing intensity.

3.2 Study area, materials and methods

3.2.1 Study area

This research was carried out in the framework of the Sino-German project “Matter fluxes in Inner Mongolia as influenced by stocking rate (MAGIM)” that investigates the problem of overgrazing with respect to soil, vegetation, hydrology and biosphere-atmosphere exchange. Field measurements were conducted on experimental areas of the Inner Mongolia Grassland Ecosystem Research Station (IMGERS, lat. 43°63', long. 116°70'), which is located in the Bayinxile League within the Autonomous Region Inner Mongolia, PR China (Fig. 3.1). The station is situated 60 km south of the city of Xilinhot. The grasslands of Inner Mongolia are part of the Eurasian steppe ecosystem. Main husbandry in the area is sheep farming. As

population growth and socio-economic changes have caused an increase in the number of livestock, the region is facing severe degradation problems due to overgrazing. From 1985 to 1999, livestock almost doubled from 618,400 to 1,133,000 (Tong et al., 2004).

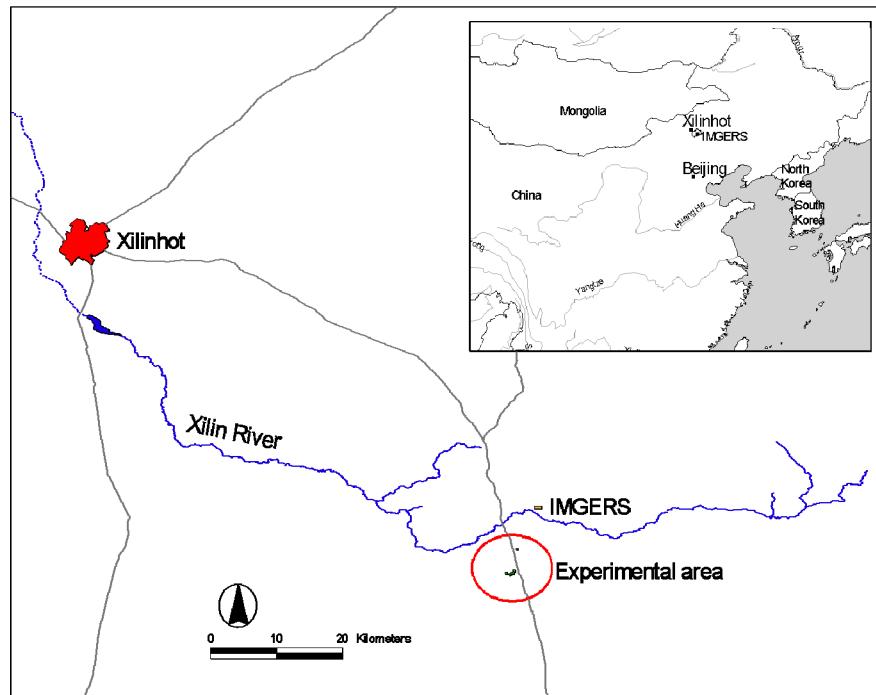


Figure 3.1. Location of the experimental area. The inset map shows the location of the study site in northern China (IMGERS: Inner Mongolia Grassland Ecosystem Research Station).

The region is marked by a continental climate with dry, cold winters and a warm, wet season in summer (average temperature in July is 18 °C, average temperature in January is -23 °C). Mean yearly precipitation is approximately 350 mm, but annual values can fluctuate between 150-500 mm. Most of the annual rainfall (60-80%) occurs in June, July and August (Chen, 1988). Nevertheless, precipitation can be highly variable during the vegetation period. Xiao et al. (1995) found coefficients of variation of approximately 50% for the summer months.

The potential climax vegetation consists of steppe grass species, such as *Leymus chinensis* and *Stipa grandis*. On degraded areas, herbaceous species, such as *Artemisia frigida*, *Cleistogene squarrosa* and *Potentilla acaulis*, are indicator species for overgrazing (Tong et al., 2004). Ongoing degradation due to heavy grazing in our study area has been detected by Tong et al. (2004). They developed a steppe degradation index (SDI) derived from satellite imagery combined with field surveys to detect the degree and spatial extent of degradation. They reported an increase of degraded area from 7,191 km² in 1985 to 7,689 km² in 1999, which corresponds to a change of approximately 4% of the catchment area.

The experimental areas of IMGERS comprise several areas that represent distinct stages of steppe vegetation as well as of (over-) grazing. Two areas are fenced and have been protected from grazing since 1979 (ungrazed 1979) and 1999 (ungrazed 1999). Grazing of different intensities can be found on three additional plots and increases gradually from winter grazing to continuous grazing and heavy grazing. The winter grazing site is enclosed during the vegetation period and has a grazing intensity of 1 sheep/ha. The continuous grazing and heavy grazing sites are browsed the whole year round with grazing intensity of 1.6 sheep/ha (continuous grazing) and 4.0 sheep/ha (heavy grazing).

The ungrazed 1979, ungrazed 1999, winter grazing and continuous grazing sites are dominated by *Stipa grandis* and *Leymus chinensis* steppe vegetation, whereas the grazing impact on the heavy grazing site is indicated by a shift in vegetation composition towards herbaceous species such as *Artemisia frigida* and *Potentilla acaulis* (Giese, pers. communication). The plant composition in each treatment itself is homogenous except for the ungrazed 1979 site. Here, the vegetation appears patchy with a mix of *Stipa grandis*, *Leymus chinensis* and *Caragana microphylla* bushes. If litter material is also considered, the heavy grazing treatment has the lowest and the ungrazed 1979 treatment has the highest soil cover. The heavy grazing site is the only treatment where bare soil is visible. The soil of all other treatments is covered with plant or litter material.

Soil texture of the experimental areas was analysed by Hoffmann et al. (2008). Results indicate similar grain size distributions on all five sites, with sand fractions ranging from 48% to 51%, silt fractions ranging from 32% to 36% and clay fractions ranging from 16% to 17%. Krümmelbein et al. (2006) and Steffens et al. (2008) analysed bulk density along vertical soil profiles and for the geostatistical grids, respectively. In the top 4-8 cm, bulk density was lowest on the two ungrazed sites and increased with grazing intensity. Topography on the five sites is uniform, and no mounds or depressions are present. Soil characteristics and a summary of the topographic analysis by Hoffmann et al. (2008) are given in Table 3.1.

Table 3.1. Grain-size distribution and topographic characteristics on the five grazing treatments after Hoffmann et al. (2008) and Steffens et al. (2008) (OC: organic carbon; BD: bulk density).

Treatment	Sand [%]	Silt [%]	Clay [%]	BD [g/cm ³]	OC [mg/g]	Slope [%]	Aspect	Topography
Ungrazed 1999 (n = 44)	47.8	36.2	16.1	1.09	25.5	5.4	N-NE	gentle and uniform slope
Ungrazed 1979 (n = 86)	49.9	34.4	15.9	0.94	31.0	4.8	NE-N	gentle and uniform slope
Winter grazing (n = 10)	48.4	34.2	17.4	1.09	25.9	5.4	N-NE	gentle and uniform slope
Continuous grazing (n = 18)	48.1	34.7	17.2	1.17	23.0	2.7	E-SE	flat
Heavy grazing (n = 80)	51.0	32.1	16.8	1.28	17.0	2.1	E-SE	flat

3.2.2 Sampling setup

Spatial and temporal trends of soil moisture distribution were compared for the five treatments during a drying period in summer 2005. The preceding period had almost no rainfall and the soil was very dry in the beginning of June. After a precipitation event from June 13th to 15th, soil moisture was monitored daily from June 16th to 22nd. The weather was warm and dry with the exception of a small rainfall event on June 19th, providing reasonable conditions to follow the drying up of the topsoil.

The soil moisture measurements were made on a rectangular sampling grid, which was set up on each of the five treatments. The sampling distance was 15 m on the regular grid and 5 m in the nested areas (Fig. 3.2). Overall, 100 sampling points were measured at each site. With this setup, the correlation of data pairs in several distance classes can be captured. The nested sampling was used to capture the small-scale variability.

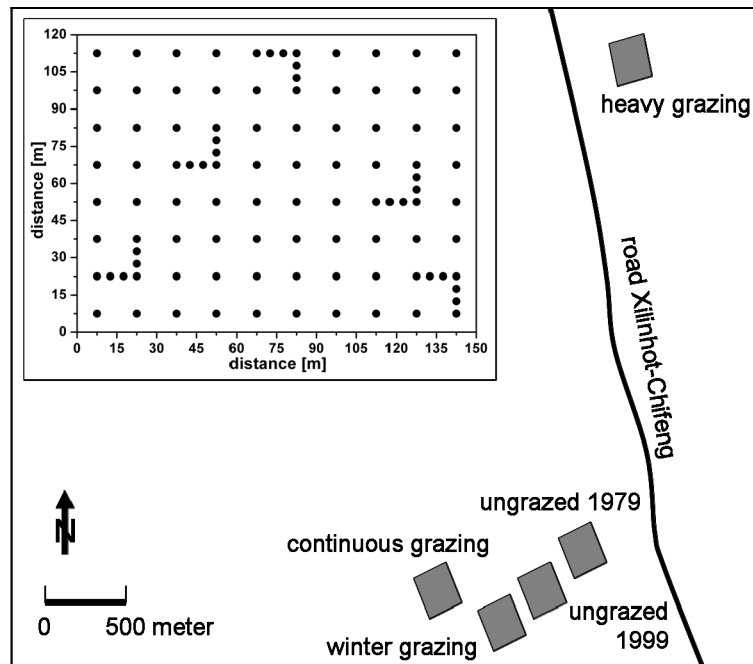


Figure 3.2. Location of the five grazing treatments. The inset map shows the sampling setup on each treatment.

We used capacitance sensors (ML2x Theta probe, Delta T Devices Ltd., Cambridge UK) for measuring the soil moisture. The sensor rods of the Theta probe are 0.06 m in length and have a diameter of 0.025 m. The probe measures the integrated moisture content in the cylindrical volume spanned by the rods. The rods were inserted vertically into the topsoil. Soil moisture was obtained from the measured signal with the built-in calibration curve, which fits well to independent moisture measurements that were conducted in the field.

3.2.3 Geostatistical analysis

The geostatistical analysis was carried out with the free software R (R Development Core Team, 2004) and the implemented package GSTAT (Pebesma and Wesseling, 1998). The normality of the distribution of each data set was checked with a Shapiro-Wilk test. If the data were not normally distributed, we searched for outliers defined as values above the 99.0 or below the 1.0 percentile. As outliers create artefacts during semivariance calculation and kriging interpolation, we removed one or two values per day and treatment in general. This led to normal distribution of all soil moisture data sets.

The geostatistical analysis investigates whether sampling points close to each other are more similar than sampling points with a larger separation. The relation of two samples separated by a distance h can be expressed by the experimental semivariogram $\gamma(h)$ (further referred to as the variogram):

$$\gamma(h) = \frac{1}{2N} \sum \{z(x_i) - z(x_i + h)\}^2 \quad (3)$$

where N denotes the number of data pairs in a particular distance class (or lag) h , $z(x_i)$ is a sample z located at x_i , and $z(x_i + h)$ is a sample separated from $z(x_i)$ by the distance h . Due to the sampling scheme, there are different distance classes h .

A variogram model is fitted to the experimental variogram to interpolate the data to unsampled locations. The spatial relationship of the measurements can be explored with three components of the variogram function: nugget, sill and range. The nugget marks the initial semivariance and indicates variability at distances smaller than the shortest sampling distance. The difference between the nugget and the highest values of $\gamma(h)$ is called the sill. Without the presence of a nugget variance, the sill is the overall span of the semivariance within the population. The range indicates the separation where the semivariance reaches its maximum, which equals the separation below which measurements are correlated.

The most common variogram functions are bounded models with a fixed range (i.e. circular or spherical models), or models that approach the sill asymptotically (exponential or Gaussian models). In our study, we chose the spherical model as it fit the data best. It was fitted with a weighted least square approach (Pebesma, 2000). The weight w for each point in the experimental variogram was defined as $w = N_i / \gamma(h_i)^2$, which relates the numbers of pairs to their semivariance at a certain lag. This weighting option strongly weights low $\gamma(h)$. If we assume that variation (and thus the variogram) is at a minimum when points are located close to each other we would expect that $\gamma(h)$ approaches 0 for small separations. However, measurements show that this is not necessarily the case. Variance at or close to the origin can be due to sampling errors, inaccuracy of the sampling device, or spatial variability below the shortest sampling interval chosen (Ancill et al., 2002). The small-scale spatial variability in particular can be considerable, because soil moisture is a highly variable parameter that is affected by other parameters subject to spatial variation, like soil texture, bulk density, plant composition or transpiration and evaporation.

The variogram model can be used to estimate soil moisture at unsampled locations via kriging. Unless a drift factor has to be considered, ordinary kriging is appropriate to predict the spatial distribution of a single variable. For more

information on geostatistical analysis, the reader is referred to standard textbooks (e.g. (Webster and Oliver, 2001)).

3.3 Results and discussion

3.3.1 Top soil water content during a wetting-drying cycle

The daily precipitation and the soil moisture development are shown in Fig. 3.3 for each experimental field. Long term rainfall data from IMGERS and additional data from rainfall loggers installed in the Xilin river basin from 2004-2006 reveal that the period presented is representative for a typical wetting and drying situation. A frequency analysis of rainfall data for the vegetation period of 2004, 2005 and 2006 showed that events with up to 2 mm of rainfall are most common in this area, and events with approximately 10 mm of rainfall occur 5 to 10 times per vegetation period (see inlet in Fig. 3.3). We further tested available rainfall data for their spatial representativeness, because the heavy grazing site is located 3 km away from the IMGERS, and hence, might experience different local rainfall patterns. We compared measured precipitation at the winter grazing site with measurements conducted at IMGERS. The two data sets show a correlation coefficient of $r=0.98$ during the studied period. Although rainfall at IMGERS was slightly lower, we consider it unlikely that the soil moisture measurements are affected by spatial variability of rainfall because the difference in observed rainfall was small and in the range that might be expected when different instruments are used.

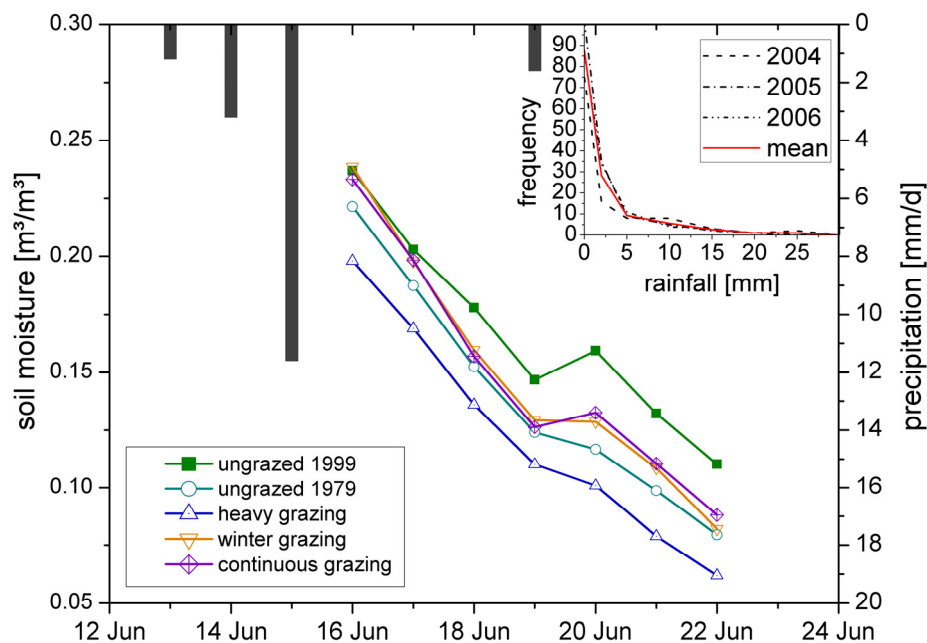


Figure 3.3. Mean ($n=100$) soil moisture and daily sums of precipitation during the sampling period in 2005. Inlet shows frequency distribution of rainfall events for 2004, 2005, 2006 and the 3-year average.

The mean soil moisture was highest at the beginning of the measuring period, shortly after the sequence of rainfall between June 13th and 15th. Prior to that rainfall event, soil moisture ranged between 0.04 [m³/m³] (ungrazed 1999, heavy grazing, continuous grazing and winter grazing sites) and 0.07 [m³/m³] (ungrazed 1979 site). After the rainfall event from June 13th to 15th (total amount 16 mm, 11.6 mm on June 15th only) soil moisture rose to 0.20-0.24 [m³/m³] on all fields. Despite this relative high amount of rainfall no surface runoff or ponding was observed across the sites. Table 3.2 outlines the sample statistics for the measured period and the results of a Tukey HSD (honestly significantly different) test comparing differences of the soil moisture means at a 0.05 significance level. This test allows the paired comparison of differences between the means of multiple data sets and calculates which data sets significantly differ from each other and which do not. The heavy grazing site displayed the lowest and the ungrazed 1999 site the highest mean soil moisture values throughout the whole sampling period. Overall, the ungrazed 1999 treatment was significantly different from the heavy grazing, but also from the ungrazed 1979 site on all sampling dates. In addition, during the last five sampling days, the ungrazed 1999 site was also significantly different from the winter and continuous grazing sites.

Table 3.2. Mean and standard deviation (SD) of soil moisture measurements on the five grazing sites.

Date	Ungrazed 1999		Ungrazed 1979		Winter grazing		Continuous grazing		Heavy grazing	
	Mean	SD	Mean	SD	Mean	SD	Mean	SD	Mean	SD
08.06.2005*	0.046 ^a	0.018	0.073 ^b	0.017	0.041 ^{cd}	0.011	0.046 ^{ad}	0.015	0.043 ^{ad}	0.011
16.06.2005	0.237 ^a	0.020	0.221 ^b	0.027	0.239 ^a	0.015	0.233 ^a	0.019	0.198 ^c	0.024
17.06.2005	0.203 ^a	0.021	0.187 ^b	0.025	0.198 ^a	0.018	0.198 ^a	0.019	0.169 ^c	0.021
18.06.2005	0.178 ^a	0.021	0.152 ^b	0.024	0.160 ^b	0.015	0.157 ^b	0.018	0.136 ^c	0.019
19.06.2005	0.147 ^a	0.021	0.124 ^b	0.026	0.129 ^b	0.013	0.126 ^b	0.017	0.110 ^c	0.018
20.06.2005	0.159 ^a	0.026	0.117 ^b	0.022	0.129 ^c	0.013	0.133 ^c	0.018	0.101 ^d	0.015
21.06.2005	0.132 ^a	0.020	0.099 ^b	0.019	0.109 ^c	0.014	0.111 ^c	0.016	0.079 ^d	0.014
22.06.2005	0.110 ^a	0.020	0.079 ^b	0.018	0.082 ^{bc}	0.011	0.088 ^c	0.015	0.062 ^d	0.013

Superscript letters (a, b, c) indicate significant ($p < 0.05$) differences of the means between sites for each sampling date.

*Due to rounding of the last digit, significant difference exists between ungrazed 1999 and winter grazing, but not between continuous grazing and winter grazing.

To explain the different reactions to precipitation of the two ungrazed sites, the management history and the vegetation composition effects on the interception storage capacity should be considered. The vegetation on the two ungrazed sites differs strongly. A mixture of grass and bushy vegetation (e.g. *Caragana mycrophylla*) grows on the ungrazed 1979 site, whereas the ungrazed 1999 site is

more homogenous and consists mainly of grasses. The different types of vegetation could lead to different infiltration, drying and transpiration mechanisms. On the ungrazed 1999 site, precipitation is not intercepted by the comparably large leaf area of shrubs and can infiltrate into the soil, but the litter cover reduces soil evaporation. The lush structure of the litter acts like a capillary barrier, allowing infiltration but no reverse capillary rise of water as was shown by Yanful et al. (2003). Litter cover on the ungrazed 1979 site is comparable, and the two ungrazed sites adjoin and have comparable exposition and slope. Hence, it is likely that the bushy vegetation on the 1979 site intercepts more rainfall, leading to significantly different soil water contents on the two ungrazed sites.

The tendency of wetting and drying is the same for all fields, but drying rates differ among sites. For example, on the first two days of the sampling period, the ungrazed 1999, continuous grazing and winter grazing sites had a similar mean value, but the continuous and winter grazing sites dried up faster in the consecutive days. After the small precipitation event on June 19th (1.6 mm), the five sampling sites showed an interesting behaviour: while soil moisture content in the ungrazed 1999 site reacted to the rainfall and the mean value clearly rose, water content in the ungrazed 1979 site and the heavy grazing site continued to decrease, although at a reduced rate. Again, continuous and winter grazing showed an intermediate behaviour. The observation is surprising as the ungrazed 1999 site and the ungrazed 1979 site are adjacent areas, so that spatial variability of rainfall is unlikely to be the reason for the observed differences. A possible explanation for the different wetting and drying rates is the degree of soil cover on the five sites. The soil on the two ungrazed sites is covered with a lot of litter material, but plant composition differs which is likely to cause differences in interception as described above. The continuous and winter grazing sites have less, and the overgrazed site has almost no litter coverage. This leads to distinct soil evaporation rates on the five sites, that accounts for the overall drying and wetting cycles of the treatments. The differences between the continuous grazing and winter grazing sites are not very distinct, which explains why the soil moisture values do not differ significantly on any sampling date. The heavy grazing site showed the lowest moisture values after precipitation and throughout the sampling period. Plant composition differs substantially from all other areas and soil coverage of living and dead biomass is much sparser as compared to the other sites. Evaporation is higher on this treatment, thus causing lowest soil moisture conditions across all treatments, as was confirmed in a modelling experiment by Wen et al. (2007).

To compare the heterogeneity of soil moisture inherent in each grazing treatment, we analysed the large-scale soil moisture variability of the single fields given by the sill of the variogram function. The background noise of small scale variability is excluded in this comparison. The box-plots in Fig. 3.4 show the distribution of the sill values during the sampling period for each treatment. The span of sill values was highest for the ungrazed 1979 treatment, illustrating the large-scale variability of soil moisture in this dataset. The winter grazing and continuous grazing sites showed less large-scale variation. Although there seems to be considerable variation in sill variance between the treatments, one has to consider the absolute differences in soil moisture, which are only 0.03-0.05 [m³/m³] for the single sampling dates (Fig. 3.3). These inter-site differences are in the same range as the day-to-day changes on the individual treatments. When comparing absolute sill values (i.e. nugget plus sill), the features shown in Fig. 3.4 remain the same, although with increased minimum values (data not shown). Only on the ungrazed 1979 site, the difference between maximum

and minimum values declines, indicating that micro-scale variability seems to be an important component of soil moisture distribution on this site.

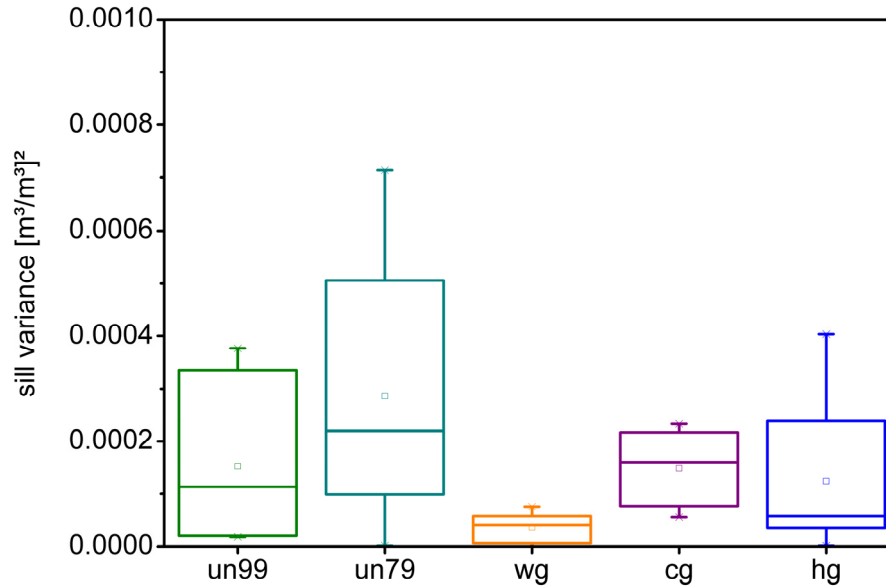


Figure 3.4. Box-plots showing the range of sill variance during the sampling period for each grazing treatment.

In order to investigate the effects of wetting and drying on soil moisture variation, we compared the semivariance for three selected sampling days: June 16th (after the major precipitation event, wet state), June 20th (after the minor precipitation event) and June 22nd (dry state) (Fig. 3.5). Calculation of directional variograms did not show anisotropy except for one day at the ungrazed 1999 site. We therefore conclude that the use of omnidirectional variograms is appropriate and that spatial trends can be neglected in the context of this work. The winter grazing and continuous grazing sites in the wet, medium and dry states have a low overall semivariance, which slightly decreases during drying. Similarly, the sill variance decreases from the wet to the dry state on the heavily grazed site, but spatial variation is higher than on the other grazed sites. The ungrazed 1979 site reaches the highest sill on June 16th and the semivariance remains comparably high, whereas the ungrazed 1999 site does not show organised spatial variation in the wet stage, a best fit to the data is accomplished by using a nugget variogram. However, after the minor rainfall on June 19th, a low sill is visible on June 20th on the ungrazed 1999 site. In general, the semivariance slightly decreases with ongoing drying for all investigated treatments. This finding is in agreement with other authors who observed increasing variance with increasing wetness (Famiglietti et al., 1998, Parent et al., 2006).

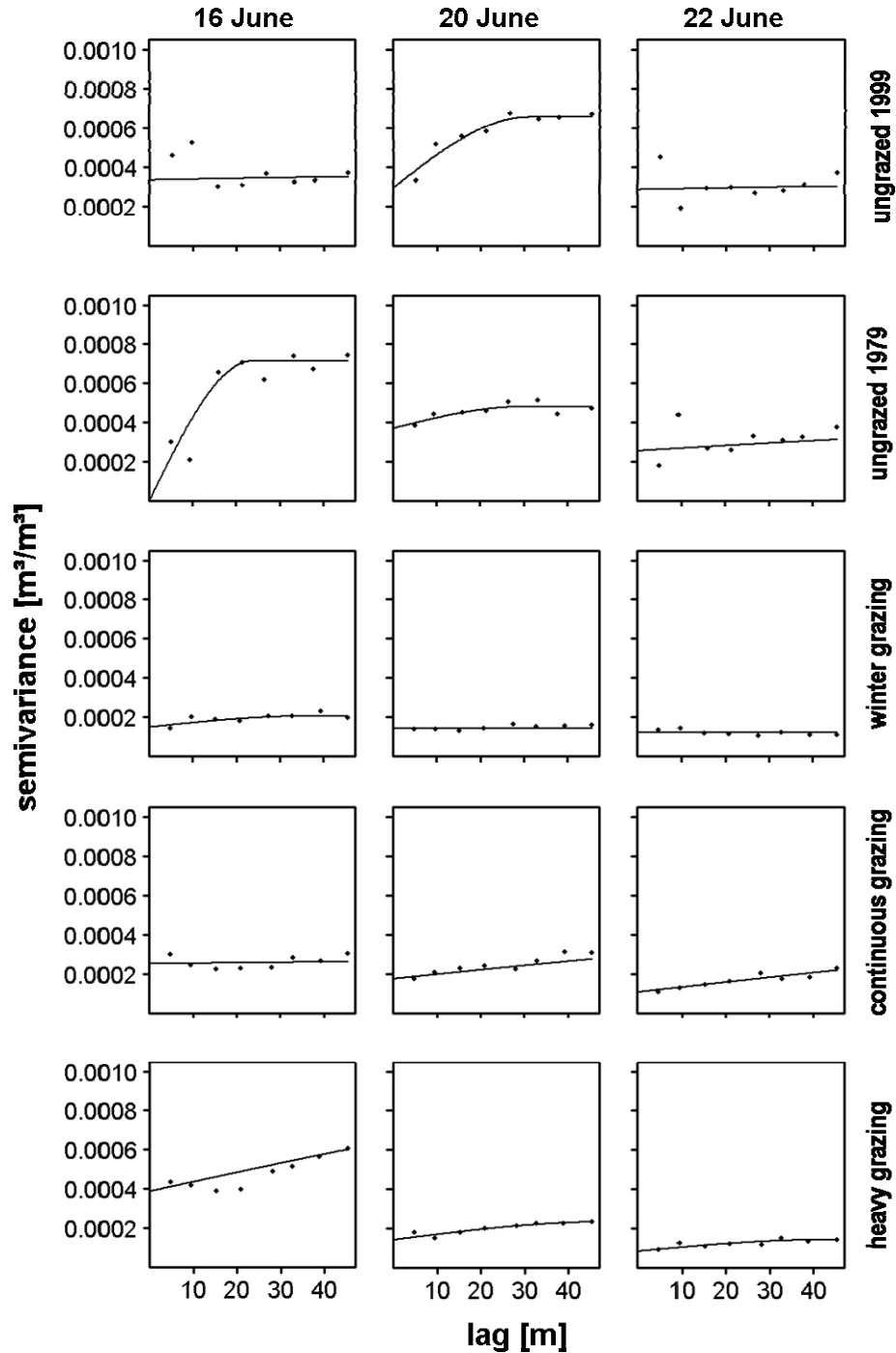


Figure 3.5. Experimental (dots) and modelled (lines) variogram of all treatments for selected days.

The fitted variogram functions were used to interpolate maps for the five grazing treatments for the period June 16th-22nd (Fig. 3.6). These maps confirm the preliminary examination of wetting and drying cycles as displayed in Fig. 3.3. The interpolation of the point measurements reveals spatial patterns in some cases, though variation is not very pronounced. Apart from a general homogenous appearance of soil moisture distribution across all sites, a distinctly different pattern is

shown for the ungrazed 1979 site (Fig. 3.6). On June 16th, dry patches of soil moisture values around 0.15 [m³/m³] alternate with wet patches of up to 0.30 [m³/m³], representing the highest soil moisture values recorded during the period of investigation. On some sites, slightly wetter areas can be tracked throughout the sampling period. Such features were observed in the eastern part of the heavily grazed site and in the south-western part of the continuous grazing site. Drying of the topsoil does not change these spatial patterns, indicating that the patterns are systematic and not due to sampling errors. The secondary rainfall event on June 19th that affected the mean soil moisture content also had an effect on the spatial variability, as shown in Fig. 3.6. One could argue that other sites, too, display patches and spots that point to spatial heterogeneity (Fig. 3.6, ungrazed 1999 site on June 17th, ungrazed 1979 site on June 18th and 19th, winter grazing site on June 18th). These features appear when the range of the variogram function is shorter than the 15 meter separation of the regular grid. This causes measurements to be singled out during interpolation. This effect is partly due to the weighting scheme applied to fit the variogram model to the experimental variogram, as previously discussed.

The general response of the five sites can be summarised as follows: according to analysis of the sill (Fig. 3.4), the variogram (Fig. 3.5) and the interpolated maps (Fig. 3.6), the ungrazed 1979 site displays the highest spatial variation of soil moisture. The other sites exhibit lower spatial variation during both, wetting and drying. With the exception of the ungrazed 1979 site, the areas exhibit a rather homogenous vegetation cover. Gómez-Plaza et al. (2001) and Cantón et al. (2004) explained the differences in soil moisture distribution across various semi-arid areas of Spain by the occurrence of different plant rooting systems that create preferred pathways for infiltration water into deeper soil layers. *Caragana microphylla* shrub vegetation in the ungrazed 1979 treatment might create similar preferred pathways for infiltration.

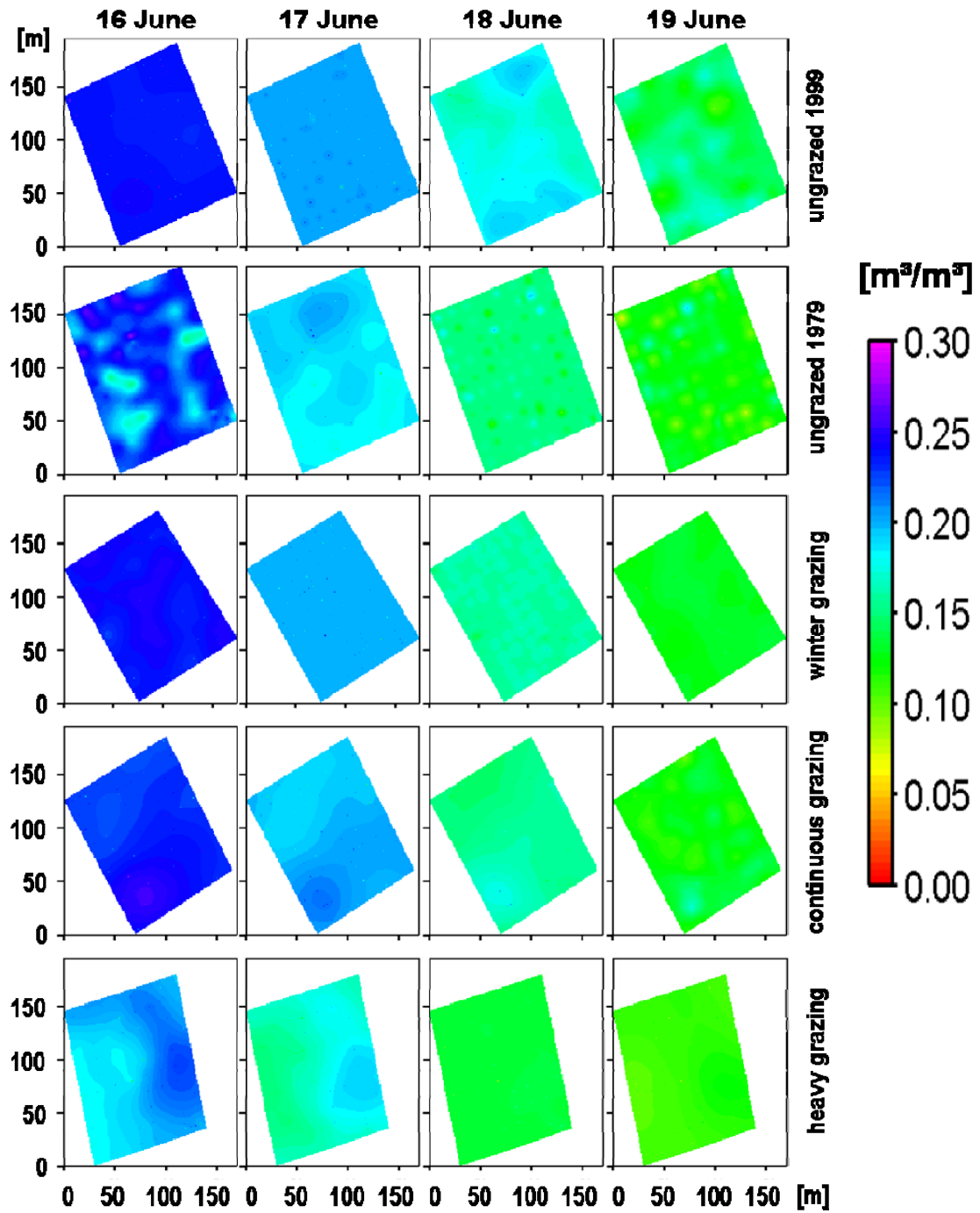


Figure 3.6. Kriged soil moisture maps for the five grazing treatments (ungrazed 1999, ungrazed 1979, winter grazing, continuous grazing and heavy grazing), 16–22 June 2005.

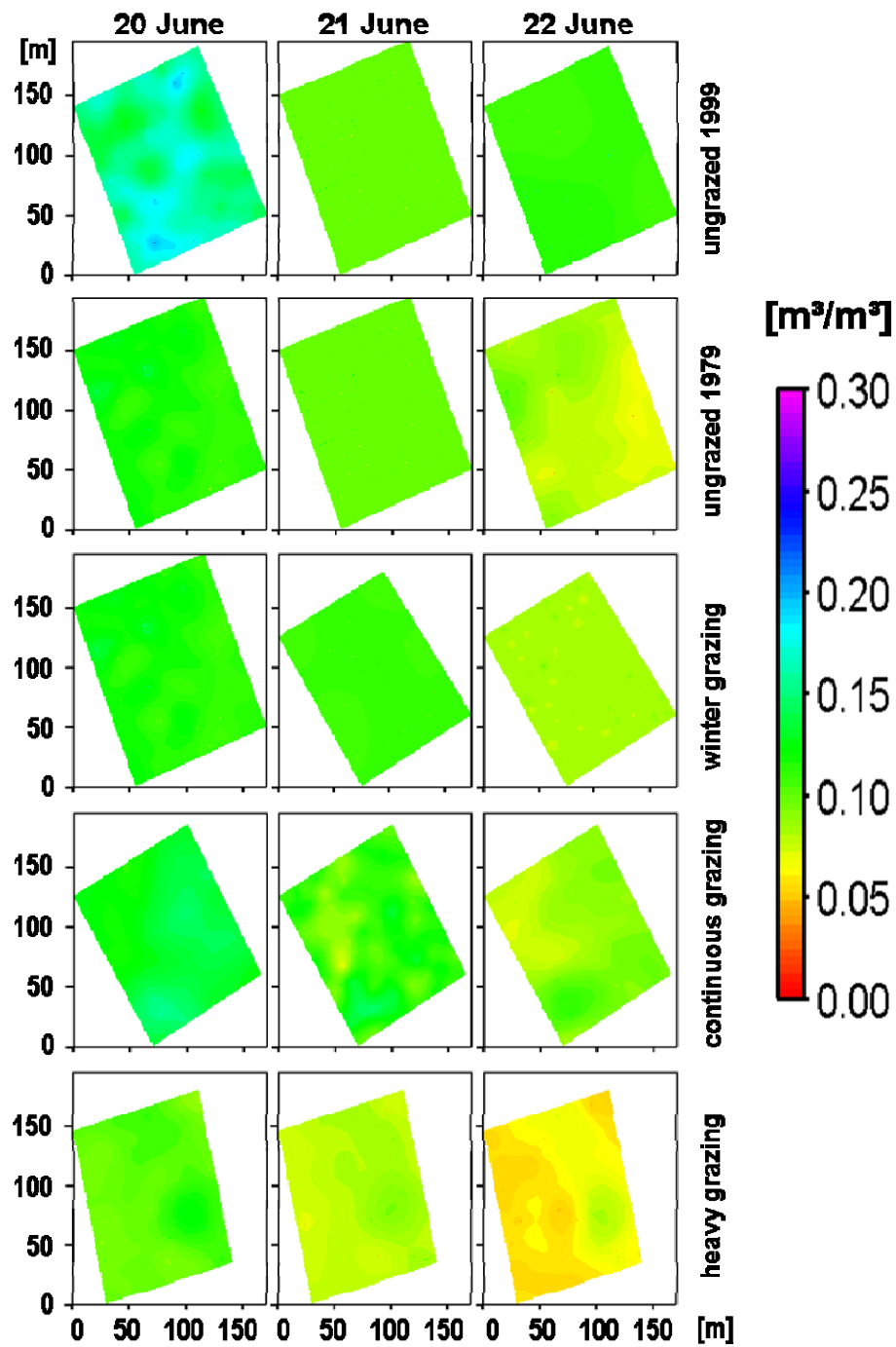


Figure 3.6. (Continued)

3.3.2 Understanding the influence of grazing on water fluxes

Recalling the initial hypothesis of this study, we found that even though differences in mean soil moisture occur across the treatments, these differences are rather small (Fig. 3.3, Table 3.2). The five sites did reveal some spatial patterns, but these were not very distinct. The capability of retaining water in the topsoil declined along the grazing gradient from ungrazed to heavily grazed. However, this conclusion is only partly valid, as the two ungrazed sites (ungrazed since 1979 and ungrazed since 1999) reacted differently to wetting and drying and showed significant differences on all sampling days. We therefore conclude that an ambiguous rather than a linear impact of grazing intensity on soil moisture content exists.

One could argue that differences of the water content in the top soil are due to other soil properties than the ones presented in Table 3.1, such as repellency or percolation rates into deeper soil layers. Zhao et al. (2007) investigated water drop penetration times across the sites for different soil water statuses as an indicator for water repellency. They stated that differences in the accumulation of organic matter between the grazing intensities might lead to differences in water repellency. However, a closer look at their data revealed that no clear pattern between the grazing intensities occurred. Measurements of soil water content along vertical profiles indicated that percolation is restricted to large rainfall events which are less frequent than the event shown in this study (Zhao, Peth and Horn, pers. communication). While soil moisture content in the upper 0.2 m reacted slightly to such rainfall events, no reaction was detected at 0.4 m. More frequent rainfall events with up to 10-15 mm as analysed in this study did not affect soil moisture in deeper soil layers at all.

Based on our own measurements and available information from other researchers working on this site, we present our conceptual understanding of vertical hydrological fluxes in this steppe environment in Fig. 3.7.

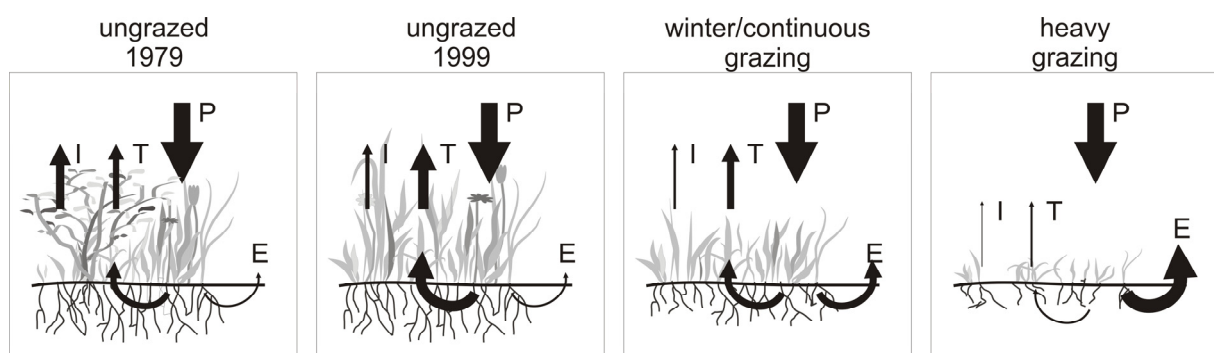


Figure 3.7. Conceptual model of plot scale vertical hydrological fluxes at natural and overgrazed sites of Inner Mongolian steppe ecosystems (P, precipitation; E, evaporation; I, interception; T, transpiration).

Dense vegetation and litter cover on the ungrazed 1979 and ungrazed 1999 sites cause high transpiration, but low evaporation rates. Fan et al. (2008) measured plant

cover, plant height and LAI of living biomass in 2005 and 2006. They relate their measurements to the NDVI (Normalized Difference Vegetation Index) of the ungrazed 1979, winter grazing and heavy grazing sites, indicating that LAI is highest on the ungrazed 1979 and lowest on the heavy grazing site. It is therefore likely that interception losses are higher on the ungrazed than on the grazed sites. Due to the canopy structure of the ungrazed 1979 site with its substantial share of shrubs, interception losses at this site are even higher than at the ungrazed 1999 treatment. As grazing impact increases and vegetation cover and LAI decreases, i.e. on the winter and continuous grazing sites, interception losses are reduced and evaporation and transpiration rates are balanced out. Finally, the large proportion of bare soil on the heavy grazing site leads to high evaporation rates that compensate for low transpiration and interception losses. This conceptual partitioning of water fluxes is backed by a modelling study of Wen et al. (2007). They simulated atmospheric water fluxes using a plot scale water and energy balance model at the same location. The results indicate that the partitioning of evaporation and transpiration is affected by grazing, but the sum of the two components does not change across the grazing sites. This is in accordance with measurements performed by Song (1996) in an adjacent *Leymus chinensis* steppe grazing gradient and a modelling study by Chen et al. (2007) in a Mongolian grassland ecosystem. We therefore conclude that most of the throughfall is quickly transpired or evaporated. This is further confirmed by analyses of measured (Ketzer et al., 2008) and modelled evapotranspiration rates (Schneider et al., 2007) at the same experimental sites. Using the eddy covariance technique, Ketzer et al. (2008) showed that water fluxes from the ungrazed and the heavily grazed site were very similar and showed no significant differences between treatments over a six weeks summer period in 2005.

3.4 Conclusions

Comparing the geostatistical results of all treatments, we find that the temporal dynamics of soil moisture is higher than the spatial variation within and between the treatments. Two reasons can explain this finding. Firstly, precipitation and dry periods control wetting and drying of the fields, rather than soil properties, grazing impact, the influence of vegetation by redistributing water, and water consumption by different plant species. Even though spatial heterogeneity of soil moisture was more pronounced on some treatments than on others, we conclude that differences within and between the treatments are rather small in this steppe environment. Secondly, as none of the fields are characterized by mounds, depressions or slopes, topographic effects on soil moisture distribution can be excluded.

The assumption that grazing intensity influences water fluxes is ambiguous. By only comparing the two ends of grazing intensity, the ungrazed 1979 and the heavily grazed site, we would not be able to see grazing effects, pointing to the fact that investigations of gradients rather than a simple paired site approach helps to understand ecosystem functioning. We acknowledge the possibility of differences in the mechanisms of plant water uptake and different rooting depths on the five sites. Yet, the results of this and further work in the study area underline the importance of evapotranspiration during the summer months which seems to superimpose the effects of grazing on water fluxes.

The lack of a clear impact of grazing intensity on soil moisture content and distribution might also be due to the investigated gradient and its management history. We do not have information about land management and use of the sites before 1979. On the one hand, one could argue that a documented management period of 26 years is a substantial period of time to result in differences due to grazing. On the other hand, there are a number of studies that report that only very long time spans of differences in grazing management lead to significant effects, such as changes in plant density and composition, reduced biomass production or decrease in basal area (Fernandez-Gimenez and Allen-Diaz, 1999; Fuhlendorf et al., 2001; Briske et al., 2003).

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4 Temporal stability of soil moisture in various semi-arid steppe ecosystems and its application in remote sensing

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Summary

Monitoring soil moisture is often necessary in hydrological studies on various scales. One of the challenges is to determine the mean soil moisture of large areas with minimum labour and costs. The aim of this study is to test temporal persistence of sample locations to decrease the number of samples required to make reliable estimates of mean moisture content in the top soil. Soil moisture data on four experimental sites were collected during the vegetation period in 2004, 2005 and 2006. The experimental sites are located in a steppe environment in northern China, and are characterised by different grazing management which causes differences in vegetation cover. A total of 100 sampling points per site were ranked with respect to their difference to field mean soil moisture using the time-stability concept. We tested whether (a) representative sample locations exist that predict field mean soil moisture to an acceptable degree, and (b) these locations are time-stable beyond a single vegetation period. Time-stable locations with a low deviation from mean field soil moisture and low standard deviation were identified for each site. Although the time-stability characteristics of some points varied between years, the selected points were appropriate to predict mean soil moisture of the sites for multiple years. On the field scale, time-stability and the persistence of patterns were analysed by the use of a Spearman rank correlation. The analysis showed that persistence depended on grazing management and the related plant cover. It is concluded that the time-stability concept provides useful information for the validation of hydrological or remote sensing models, or for the upscaling of soil moisture information to larger scales. A preliminary comparison of soil moisture measurements derived from ground-truth and remote sensing data showed that the data matched well in some cases, but that the considerable difference in spatial extent promotes differences in other cases.

Keywords: soil water content, time-stability, upscaling, remote sensing

4.1 Introduction

Reliable estimates of soil moisture recharge and drying are an essential prerequisite when addressing ecohydrological issues. For example, the water availability for plant growth or the watershed-scale soil moisture status are required in hydrological studies. Thus, soil moisture monitoring has been, and still is, an important issue in ecohydrological research. At the small scale, spatial and temporal soil moisture variability can be high due to the heterogeneity of soil, vegetation and topography. At larger scales, atmospheric processes superimpose moisture dynamics on small scale effects and make reliable estimates of mean soil moisture even more difficult. Therefore, a major concern during the past years has been the identification of factors controlling spatial and temporal soil moisture variability (e.g. Grayson et al., 1997; Western et al., 1999; Entin et al., 2000) and the quantification of this variability at different scales (e.g. Seyfried, 1998; Skøien et al., 2003). Famiglietti et al. (1998) were able to show that these factors not only change with scale, but that their influence also depends on the initial soil moisture condition.

According to Grayson et al. (1997), factors leading to spatial and temporal changes in soil moisture can be separated in local (e.g. soil properties and microtopography) and non-local controls (e.g. drainage lines due to catchment topography). Seyfried (1998) found that the spatial variability of soil moisture is stratified on larger scales ($> 10^8 \text{ m}^2$) depending on soil series and climatic gradients. Choi et al. (2007) compiled the results of several studies on grasslands and croplands in Europe and the USA with sampling extents ranging from field to basin scale. They applied a principal component analysis and concluded that rainfall and topography control the change in soil moisture variability under drying-wetting cycles, while soil parameters control the relative amplitude.

Capturing temporal and spatial soil moisture variability at scales from fields to watersheds can be time and labour intensive. Soil moisture estimates over large areas can be obtained with airborne or satellite-based remote sensing methods (Sano et al., 1998; Walker et al., 2004; Bosch et al., 2006). However, these methods generally have footprints of tens of meters to tens of kilometres, making it difficult to validate remote sensing data with conventional soil moisture measurement techniques. Hence, a method that can be used to estimate average soil moisture of a large area as accurate as possible, but with as little effort as necessary is appealing for validation of remotely sensed data, upscaling field experiments, or the verification of hydrological model predictions.

Vachaud et al. (1985) introduced the concept of time-stability as a tool to reduce the measurement effort to characterize large fields. In this concept, one or more locations out of a larger sample volume are selected to predict average field soil moisture. The method was developed and tested at three sites in France, Spain and Tunisia with differing climate and land cover, and with varying sample sizes. Various authors picked up this concept and included it in their soil moisture variability studies. For example, Grayson and Western (1998) identified catchment average soil moisture measurement sites in three different catchments in Australia and in the USA. In a semi-arid environment in Spain, Gómez-Plaza et al. (2000) investigated the factors controlling time-stability. The change of plant cover throughout the vegetation period proved to impact temporal persistence most. In another semi-arid catchment, Martínez-Fernández and Ceballos (2005) showed the validity of the

concept on a small (i.e. $< 1 \text{ km}^2$) and on a large scale (i.e. $> 1000 \text{ km}^2$). Kachanoski and de Jong (1988) extended the time-stability concept with spatial coherency analysis, which includes spatial aspects into the analysis of temporal persistence. In their study, spatial patterns varied between sampling dates at small scales ($< 40 \text{ m}$), but were not affected at larger scales.

Although the concept is promising, time-stability is not evident on all monitoring sites. Comegna and Basile (1994) could not identify sampling sites that were representative for the mean soil moisture. They argued that this was related to the homogeneous soil properties of the cropped sandy Andisol. In another study on a clay loam soil, Kamgar et al. (1993) found that temporal persistence of soil moisture depended on the sampling depth. Surface sampling points did not show temporal persistence, whereas soil moisture down to 2.85 m depth revealed time-stable patterns.

It is obvious that vegetation, soil properties and topography can influence the temporal persistence of soil moisture patterns. While the latter two can be considered to be constant within a certain time frame, vegetation can affect soil moisture dynamics seasonally. Especially, in water limited ecosystems like our study region, the vegetation patterns impact water storage in the soil. Comparing the time-stability of soil moisture patterns in vegetated and non-vegetated transects, Gómez-Plaza et al. (2000) showed that a transect with plant cover was much more variable in time than two non-vegetated transects. The distribution of plant cover and roots caused a varying plant water demand, which invalidated the time-stability concept during the growing season. In the dormant period, soil moisture patterns became more time persistent. In addition to vegetation abundance, temporal stability varies with wetting and drying. Martínez-Fernández and Ceballos (2003) showed that stability increased under recharge conditions and was lowest under dry conditions.

For a successful application of the time-stability concept, the selected points need to represent average moisture dynamics beyond the time period they were determined for. Martínez-Fernández and Ceballos (2003) compared the ranking of mean relative difference of the soil moisture monitoring sites in three consecutive years. Despite some fluctuations, the dry and wet sites remained stable even under different climatic boundary conditions. However, more studies are required to confirm the general applicability of the time-stability concept for multi-year datasets.

Time-stable points have been successfully used to validate remote sensing soil moisture data. In various field experiments, representative soil moisture locations across heterogeneous landscapes have been identified (Famiglietti et al., 1999; Mohanty and Skaggs, 2001; Jacobs et al., 2004; Cosh et al., 2004; Ceballos et al., 2005). Time-stable locations representing the average soil moisture of an airborne radiometer footprint were identified by Famiglietti et al. (1999) and Mohanty and Skaggs (2001). However, the degree of time-stability varied between different studies. Soil and land cover properties were identified as determining factors for time-stable behaviour. Although these applications are promising, the shortcomings and restrictions have to be addressed. While the scale of airborne and ground-based measurements matched in these studies, the remote sensing footprint of satellite imagery is typically much larger than the extent of the ground measurements, which make comparisons difficult. However, satellite remote sensing data is the only source of spatially distributed information in poorly or ungauged basins. Ecohydrological model applications often fail due to the lack of ground-based data for model validation and calibration. Remote sensing products could provide valuable input to

overcome these limitations, or provide new means for model verification. It is one of the aims of the predictions in ungauged basins initiative of the International Association of Hydrological Sciences (IAHS) (Sivapalan et al., 2003) to enhance our capabilities in catchment modelling when hydrometeorological or runoff data are missing. Thus, the benefit of soil moisture data retrieved from satellites is that it provides access to areas that are, in a hydrological context, unexplored.

In this study, we analyse the potential of time-stable points to estimate average soil water content on the field scale (approximately 1.5 ha). We aim to determine whether time-stability persists within a seasonal cycle and between consecutive growing seasons. In addition, the impact of varying vegetation structure on both point and field scale temporal stability is investigated by considering plots with different grazing management. We attempt to clarify to which degree the number of sampling points can be decreased while the accuracy of mean field soil moisture estimation remains acceptable. Sampling efforts could be considerably reduced with time-stable points and hence, time series of soil moisture could be obtained more easily. Finally, time-stable points may be an effective tool to validate remote sensing soil moisture data. Particularly in poorly gauged catchments like the study area, remote sensing data might reduce uncertainty in hydrological applications when other, ground-based data is scarce. To show the potential of the time-stability concept, we compare the average field moisture content determined from time-stable points with soil moisture data from satellite remote sensing.

4.2 Materials and Methods

4.2.1 Research area

The study was carried out in a semi-arid steppe environment in the Autonomous Province Inner Mongolia, PR China (Fig. 4.1). The region is marked by a continental climate with cold winters (average temperature in January: -23°C) and warm summers (average temperature in July: 18°C) (Chen, 1988). Mean annual precipitation is 350 mm. Most of the rainfall is received during the vegetation period with 60-80% of the annual rainfall occurring from June to August. Yet, seasonal and interannual fluctuations can be high (Xiao et al., 1995). As a consequence, high demands for plant water uptake can coincide with periods of water limitation.

The study sites are located in the Xilin river catchment (3600 km²). Soil moisture measurements were carried out on experimental areas of the Inner Mongolia Grassland Ecosystem Research Station (IMGERS, lat $43^{\circ}63'$, long. $116^{\circ}70'$) within the frame of the MAGIM Research Unit (for further details see <www.magim.net>). There are four experimental sites with different grazing intensities: winter grazing (wg, 1.0 sheep ha⁻¹), heavy grazing (hg, 4.0 sheep ha⁻¹) and two no-grazing sites, which were fenced in 1979 (ug79) and 1999 (ug99). The four sites have different vegetation composition and cover. The degree of plant and litter cover decreases with increasing grazing intensity. In 2005 and 2006, Fan et al. (2007) measured a maximum green coverage of 50% on the ug79 site in August, 40% green coverage on the wg site in July and August and 35% maximum green coverage on the hg site in July. The hg site appears to be the most homogeneous site due to constant sheep trampling, whereas the two ungrazed sites, in particular the ug79 site, show highest heterogeneity with respect to plant composition and plant distribution. The sites

represent the typical steppe environment of the Xilin river catchment. Approximately 75% of the catchment area is grassland, 6% is cropland and 16% is a sand dune belt with sparse grassland and a loose tree population. The remaining area is wetland along the Xilin river.

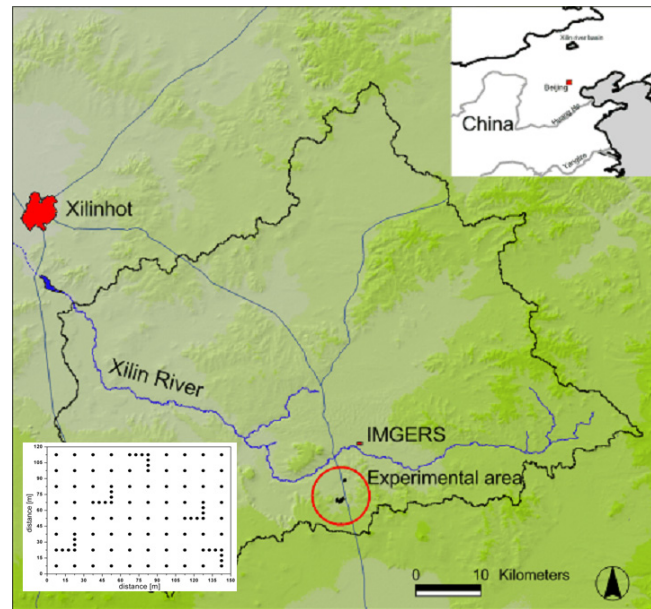


Figure 4.1. Location of the study area in northern China with an outline of the experimental sites and the geostatistical sampling grid.

Topography is undulating with a mean slope of 2.7° . Steffens et al. (2008) analysed soil texture, bulk density and organic carbon (Table 4.1). The results indicate that bulk density decreases with decreasing grazing intensity and is lowest on the ug79 site, while organic carbon increases with decreasing grazing pressure. Sand content is highest on the hg site due to wind erosion of finer particle sizes on this poorly vegetated site (Hoffmann et al., 2008).

Table 4.1. Mean values of soil characteristics in the upper 0.04 m according to Steffens et al. (2008).

	ug79 <i>n</i> =98	ug99 <i>n</i> =99	wg* <i>n</i> =122	hg <i>n</i> =98
BD [g/cm ³]	0.94 (0.10)a	1.09 (0.12)b	1.09 (0.08)b	1.28 (0.08)d
OC [mg/g]	31.00 (5.50)a	25.50 (6.3)b	25.90 (4.50)b	17.00 (4.20)d
Sand [mg/g]	49.11 (5.49)ac	46.70 (5.60)ab	43.95 (5.56)b	51.05 (5.17)c
Clay [mg/g]	16.09 (1.72)a	16.27 (2.22)a	18.30 (1.92)b	16.81 (2.09)a

Standard deviation is given in brackets. Different letters indicate significant difference at $p=0.01$. (BD = bulk density, OC = organic carbon).

*Data on wg were measured on a different grid

4.2.2 Soil moisture measurements

On each of the four sites, soil moisture of the top soil was sampled at 100 locations. The sampling grid consisted of 80 points on a regular 15 m x 15 m grid and an additional 20 points nested into the grid at 5 m separation (Fig. 4.1). Soil moisture in the upper 0.06 m was measured with a ThetaProbe (ML2x, Delta T Devices Ltd., Cambridge UK). The portable sensor uses a standing wave measurement technique. A signal is transmitted into the soil via the array formed by the four rods of the sensor. At the sensor's frequency (100 MHz), the array impedance depends mainly on the dielectric permittivity of the soil, which in turn is well correlated with soil moisture content. The output of the sensor is voltage, which was directly related to soil moisture content with the calibration relationship provided by the manufacturer. An independent evaluation of the calibration relationship with gravimetric soil samples showed that this relationship was appropriate for the soils considered in this study.

Soil moisture was measured at regular intervals during the vegetation period of 2004-2006. Time-stability was analysed for 12 sampling days in 2004, and for 8 sampling days in 2005. The remaining 2005 and 2006 data were used to validate the time-stability concept. Rainfall sums between May and September were 288 mm in 2004, 142 mm in 2005, and 263 mm in 2006. It should be noted that the soil parameters summarized in Table 4.1 were measured at the same grid locations as soil moisture at the ug79, ug99 and hg site. This allows an analysis of the relationship between time-stability and soil properties on these sites. Unfortunately, the soil properties on the wg site were measured on a different grid.

The ERS scatterometer data used in this study were provided by the Institute of Photogrammetry and Remote Sensing (IPF), Vienna University of Technology. The ERS scatterometer is an active microwave sensor with a spatial resolution of 50 km. Temporal resolution is irregular depending on the operation of the sensor. The conversion of the backscatter signal at a reference angle of 40°, $\sigma^0(40)$, into soil moisture follows a change detection method after Wagner et al. (1999). In their study, the lowest $\sigma_{dry}^0(40,t)$ and highest $\sigma_{wet}^0(40,t)$ backscatter values measured at different times t represent the range of soil moisture values. Backscatter proved to be lowest under dry conditions, and highest under wet conditions. The relative soil water content $m_s(t)$ [%], further also referred to as surface wetness index (surfwet), was derived as follows:

$$m_s(t) = \frac{\sigma^0(40,t) - \sigma_{dry}^0(40,t)}{\sigma_{wet}^0(40,t) - \sigma_{dry}^0(40,t)} \quad (4)$$

The data used in this study were preprocessed by the IPF and $m_s(t)$ was provided in percent for the top 0.05 m of the soil. Surfwet ranges from 0% (dry) to 100% (saturation), i.e. a surfwet of 0% represents wilting point and a surfwet of 100% represents total water capacity (Wagner et al., 1999). To obtain comparable data sets, we applied minimum and maximum soil moisture values measured after long drought or intensive rainfall on each site and converted the volumetric soil moisture ground measurements into a wetness index ranging from 0% to 100% moisture content accordingly. The minimum and maximum volumetric soil moisture measured between 2004 and 2006 on the four sites were 0.01 and 0.37 (ug79), 0.01 and 0.4

(ug99), 0.01 and 0.37 (wg), 0.00 and 0.33 (hg). The ground measurements provide an average of the upper 0.06 m of the soil and hence the sampling depth of ground and satellite data corresponds well.

4.2.3 Time-stability

The time-stability approach developed by Vachaud et al. (1985) assumes that average field soil moisture can be represented by single measurements selected from a larger measurement volume. The quality of time-stable points is expressed by their mean relative difference from mean soil moisture. The relative difference for a single sampling date is defined as

$$\delta_{i,j} = \frac{S_{i,j} - \bar{S}_j}{\bar{S}_j} \quad (5)$$

where $\delta_{i,j}$ is the relative difference from mean soil moisture at location i and time j , $S_{i,j}$ is the soil moisture at location i and time j and \bar{S}_j is mean soil moisture at the same time. The mean relative difference $\bar{\delta}_j$ over the entire sampling period is then calculated as

$$\bar{\delta}_j = \frac{1}{m} \sum_{i=1}^m \delta_{i,j} \quad (6)$$

with m being the number of sampling dates.

Sampling locations are considered to be time-stable when (1) they show no or only little difference from mean soil moisture over time and (2) the standard deviation of the mean relative difference is low. The second constraint is even more important as time-stable points with a low standard deviation will give more precise estimates even when their values differ from mean soil moisture. In this study, all sampling locations with a mean relative difference smaller than 0.1 and a low standard deviation were defined as time-stable.

The above method tests the temporal persistence of individual points in the field. The time-stability of the site covered by the sampling grid was estimated with a correlation analysis. We applied the non-parametric Spearman's test of rank correlation r_s :

$$r_s = 1 - \frac{6 \sum_{i=1}^n (R_{i,j} - R_{i,j'})^2}{n(n^2 - 1)} \quad (7)$$

A rank $R_{i,j}$ was assigned to the $n = 100$ soil moisture sample locations i on each site at sample date j . The change of rank at sample date j' was analysed with the

correlation of ranks. The value of r_s ranges from -1 to 1. Correlations close to 1 occur when the rank of the soil moisture samples only slightly changes between sampling dates. Thus, high correlations between sampling dates indicate temporal persistence of spatial patterns.

4.3 Results and discussion

Fig. 4.2 presents the ranked mean relative difference (and its standard deviation) for all grazing treatments for both 2004 and 2005. In general, mean relative differences deviate less than 30% from mean field soil moisture. Only the hg site has sampling locations with a higher mean relative difference. In 2004, the mean relative difference ranged between -22 and +18% on the ug79 site and between -16 and +26% on the ug99 site. Although maximum and minimum values differ on these sites, the span of mean relative difference is similar. The wg site shows a slightly lower span of $\bar{\delta}_j$ (-16 to +16%). The highest span of $\bar{\delta}_j$ occurs on the hg site (from -21 to +40%). In 2005, the shapes of the graphs changed slightly, but the range of $\bar{\delta}_j$ on all sites is similar to 2004. It should be noted that although the mean relative difference is lower than 50% on all sites, the standard deviation of $\bar{\delta}_j$ is considerable. Thus, a single measurement can deviate much more from the mean soil moisture of the field than indicated by its $\bar{\delta}_j$ value.

Precipitation in 2004 was much higher than in 2005 (288 vs. 142 mm). Therefore, we analysed the effect of a “normal” versus a “dry” year on the mean standard deviation of $\bar{\delta}_j$. We found that the standard deviation decreased on all but the hg site in 2005. Obviously, this is related to the dryer soil moisture status in 2005. Variability introduced by wet soil conditions is not present in this year, and, therefore, the standard deviation is lower.

The mean value of standard deviation slightly increased from the first half to the second half of the ranked data for the ug79 and hg sites in 2004 and 2005 and for the ug99 and wg sites in 2005. Nevertheless, there is no clear relationship between standard deviation and $\bar{\delta}_j$. The correlation between mean relative difference and standard deviation is below 0.5 on all sites. This is in contrast to Jacobs et al. (2004), who observed low standard deviations for $\bar{\delta}_j < 0$ and high standard deviations for $\bar{\delta}_j > 0$.

The observed range of mean relative difference for the four sites lies between the values found in studies with a smaller sampling extent (e.g. Vachaud et al., 1985; Comegna and Basile, 1994) and catchment scale observations with larger sampling extent (e.g. Mohanty and Skaggs, 2001; Martínez-Fernández and Ceballos, 2003). Therefore, it is tempting to conclude that the range of mean relative difference increases with scale because of an expected increase in variation of soils, topography and vegetation. However, Cosh et al. (2004) observed mean relative differences below 40% in a 100 km² watershed. Compared to studies with a similar extent and sampling depth, the values found in this study are on the low side. For example, Grayson and Western (1998) found similar values of $\bar{\delta}_j$ in a small catchment in S.E. Australia, whereas Gómez-Plaza et al. (2000) found a larger range of mean relative differences. This might partly be explained by the layout of the experiments. Grayson and Western (1998) studied several small watersheds, i.e. obtained areal information, whereas Gómez-Plaza et al. (2000) studied transects on

three slopes. The strong topographic difference along the transects might have promoted a higher range of δ_j .

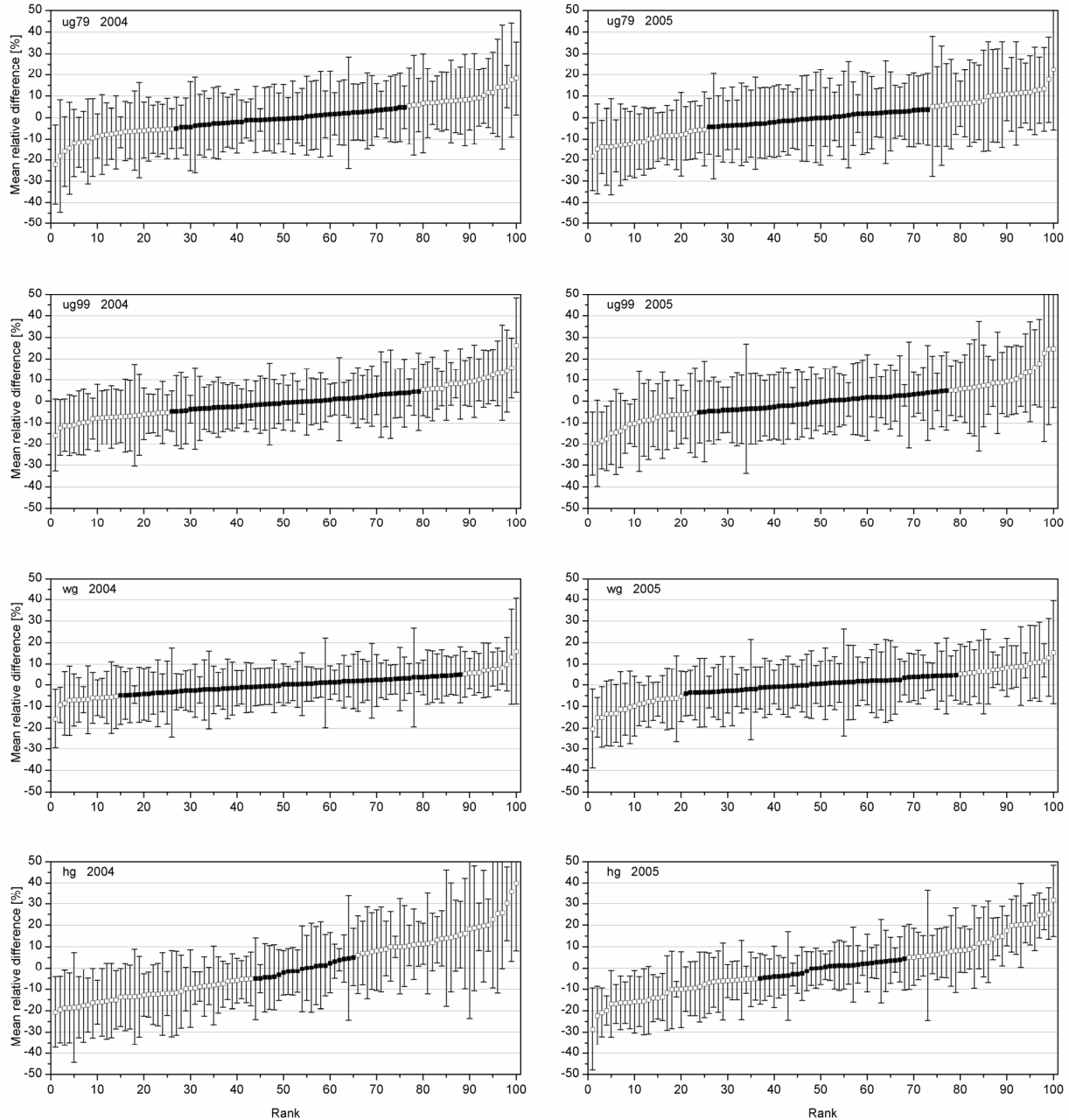


Figure 4.2. Ranked mean relative difference of soil moisture on the four grazing sites in 2004 and 2005. Sampling points with a mean relative difference below 5% are marked in black. Squares indicate the mean relative difference and the error bars one standard deviation.

Locations are considered to be time-stable when they have a low mean relative difference and a low standard deviation. Table 4.2 shows that 22–54% of the sample points are within 5% difference from the mean. In addition, more than 90% of the

sample points are located within one standard deviation from the mean at all sites except the hg site.

Table 4.2. Percentage of sample points located within 5% difference, and 1 and 2 standard deviations (std) from mean.

	ug79		ug99		wg		hg	
	2004	2005	2004	2005	2004	2005	2004	2005
5%	50	48	54	54	74	59	22	32
1std	95	97	94	92	97	93	81	70
2std	100	100	100	100	100	100	98	95

From this large population of sampling points with small relative differences, we decided that the most suitable time-stable points were those with the lowest standard deviation. These time-stable locations from 2004 were validated using the 2005 data. Fig. 4.3 shows mean soil moisture enveloped by two times the standard deviation in 2005. Soil moisture values of four time-stable points and their mean are also shown in Fig. 4.3.

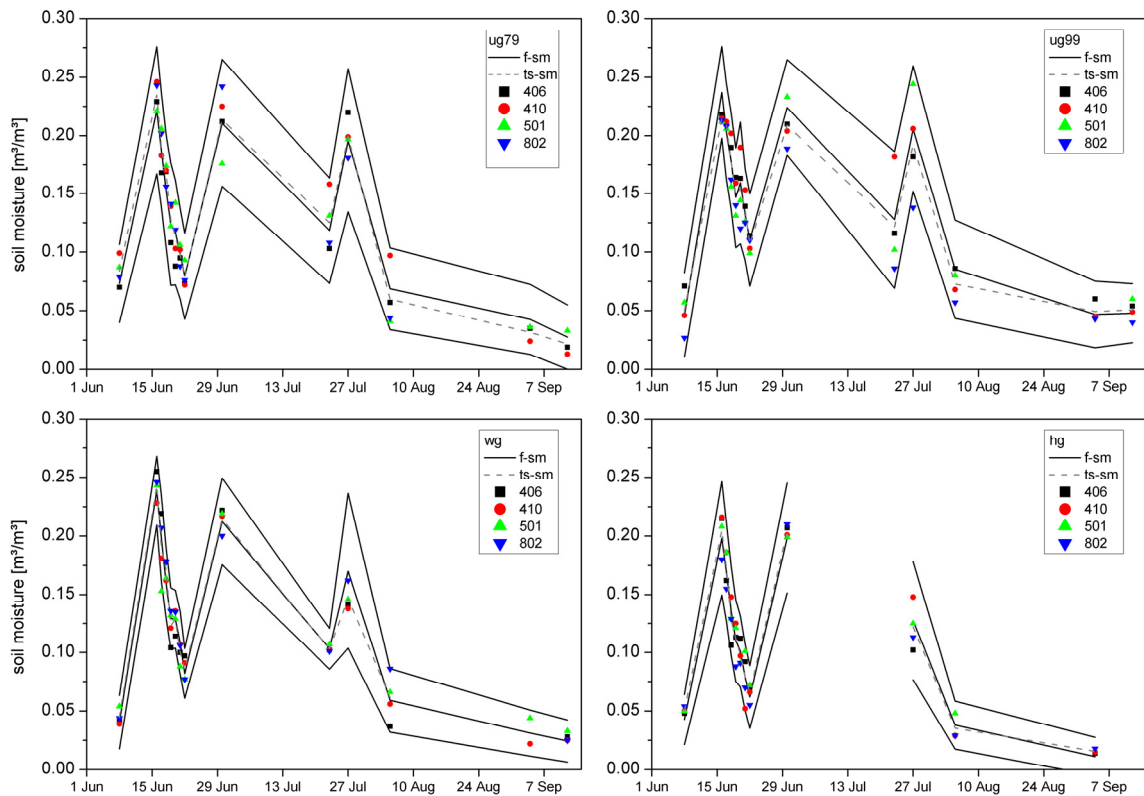


Figure 4.3. Soil moisture dynamics on the four grazing sites in 2005. Dashed line: mean soil moisture calculated from time stable points, solid line: field mean soil moisture and $\pm 2\sigma$.

Obviously, the time-stable points from 2004 provide an accurate estimate of temporal soil moisture dynamics in 2005. The time-stable points never lie outside two times the standard deviation and the mean of the time-stable points closely follows the mean field moisture content.

To test the temporal persistence of time-stable points beyond one vegetation period, we compare the quality and time-stable properties derived from the 2004 and the 2005 data set. The points selected in 2004 partly retained their time-stable characteristics in 2005, but some points were not time-stable anymore in 2005. When time-stability changed between 2004 and 2005, the points basically kept a similar mean relative difference, but showed an increased standard deviation. Nevertheless, the points selected in 2004 still performed reasonable for the 2006 soil moisture data. In Fig. 4.4, mean soil moisture in 2006 is plotted against soil moisture measured at time-stable locations that were determined from: (a) the 2004 data, (b) the 2005 data and (c) data of both years. With a few exceptions, all sets of selected time-stable points predict soil moisture within 5% error at each sampling date. This is in agreement with Jacobs et al. (2004), who limited their analysis to points with a mean relative difference below 5% and found that the five best time-stable points provide satisfactory estimates.

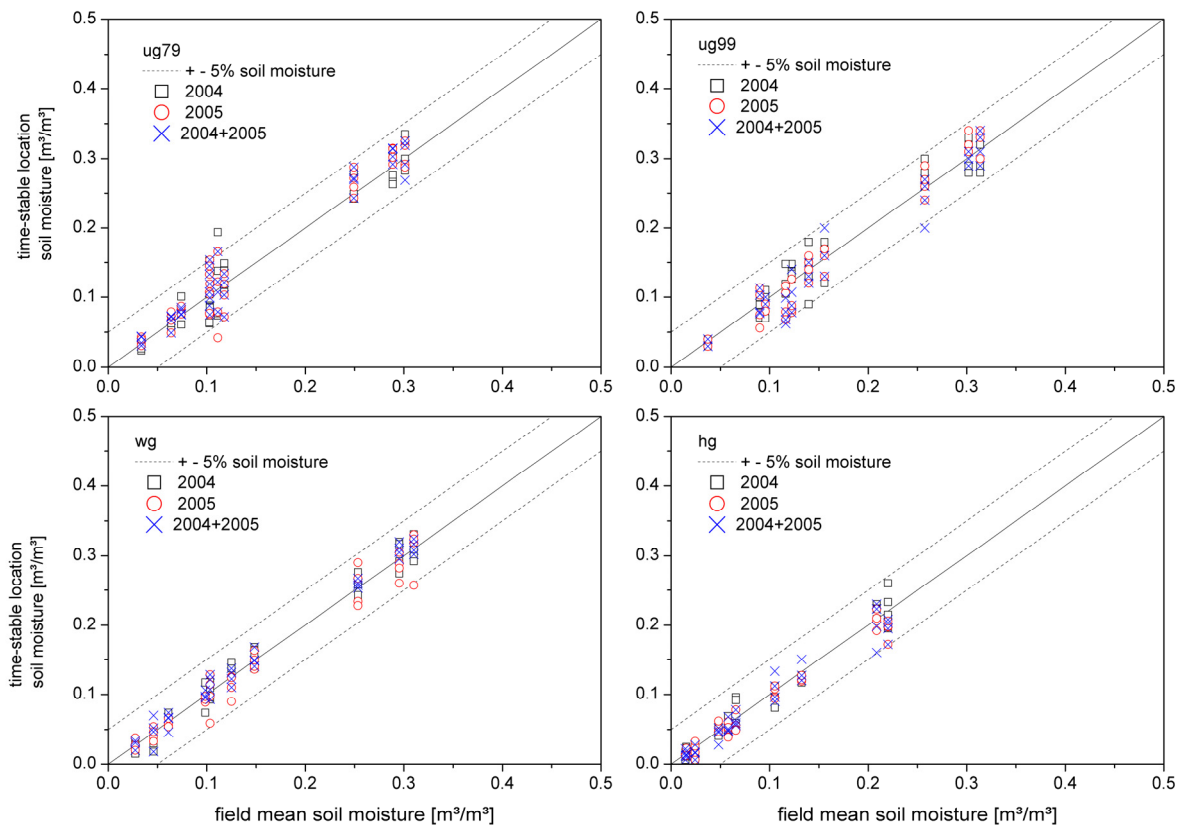


Figure 4.4. Comparison of field mean soil moisture with soil moisture of time-stable samples calculated from 2004, 2005 data and compiled from both years.

Table 4.3 presents the root mean squared error (RMSE) between mean soil moisture obtained from the selected time-stable points and the actual mean soil moisture in 2006. The RMSE is low for both the 2004 and 2005 time-stable points. In addition, combining the 2004 and 2005 soil moisture data did not improve the prediction of the 2006 data. To provide a reference for the best RMSE values presented in Table 4.3, RMSE values of the four worst (i.e. the two lowest and the two highest ranked) time-stable points in 2004 are also presented. It can be seen that the RMSE for these points is considerably higher. The 2004 and 2006 precipitation amounts are very similar, and the time-stable points selected from the 2004 data perform slightly better than the 2005 time-stable points in 2006. The much dryer conditions during the vegetation period of 2005 apparently affect the transferability of the time-stable points to wetter conditions.

Table 4.3. Mean RMSE between soil moisture of time-stable points and the 2006 mean field soil moisture.

	Best time stable points selected from 2004 data	Best time stable points selected from 2005 data	Best time stable points selected from 2004+2005 data	Worst time stable points selected from 2004 data
ug79	0.008	0.010	0.009	0.015
ug99	0.006	0.010	0.009	0.015
wg	0.006	0.006	0.004	0.024
hg	0.007	0.006	0.006	0.019

The analysis of soil moisture data from several years indicates that sampling points revealing temporal persistence throughout one season may lose their time-stable properties due to a higher standard deviation in another season, but still retain the capability to estimate average field soil moisture. The data imply that even when data of several vegetation periods are compiled, a perfect time-stable soil moisture predictor can not be obtained because the associated standard deviation will cause fluctuations around mean soil moisture. It remains open to debate whether more accurate time-stable locations with lower standard deviation can be obtained when time series longer than two years are considered. The results correspond to a study of Martínez-Fernández and Ceballos (2003). They also found that the quality of time-stable points shifted throughout the three years of their study.

To analyse whether the mean relative difference, and therewith the time-stability, depends on soil properties, a correlation analysis was performed. Table 4.4 presents the correlation between $\bar{\delta}_j$ and bulk density, organic carbon content, sand content and clay content. In contrast to Jacobs et al. (2004), the quality of time-stable points can only partly be explained by soil characteristics. On both ungrazed sites, neither bulk density, organic carbon or sand and clay content could explain the time-stable characteristics of the sampling points, as indicated by the low R^2 values (Table 4.4). Although on a weak to moderate level, correlations were consistently higher on the hg site, with sand content showing the highest correlation. Between 2004 and 2005, correlations decreased on the ug99 and hg sites, while they increased on the ug79 site albeit on a low level. This indicates that the different weather conditions in 2004 (average precipitation) and 2005 (dry) influenced the time-stability characteristics. In

contrast to Famiglietti et al. (1998), we could not identify soil characteristics with varying impact on temporal stability under dry or wet conditions.

Table 4.4. Coefficient of determination (R^2) between soil characteristics and $\bar{\delta}_j$ in 2004 and 2005.

	ug79		ug99		wg		hg	
	2004	2005	2004	2005	2004	2005	2004	2005
BD	0.001	0.022	0.171	0.142	-	-	0.279	0.239
OC	0.000	0.047	0.230	0.148	-	-	0.511	0.397
Sand	0.020	0.030	0.231	0.073	-	-	0.709	0.519
Clay	0.003	0.011	0.184	0.010	-	-	0.456	0.379

Soil properties on the soil moisture sampling locations are not available on the wg site. (BD = bulk density, OC = organic carbon)

Since the soil moisture measurements at different sampling dates are not made at exactly the same location, the minimum RMSE that can be obtained with a single time-stable point is limited by the reproducibility of the moisture measurements at each measurement location. Averaging of several time-stable points could further reduce the RMSE because the error of small scale variability at single measurement locations is averaged out. This effect is shown in Fig. 4.5. Compared to only using one time-stable point, averaging three or four time-stable points already decreases RMSE considerably.

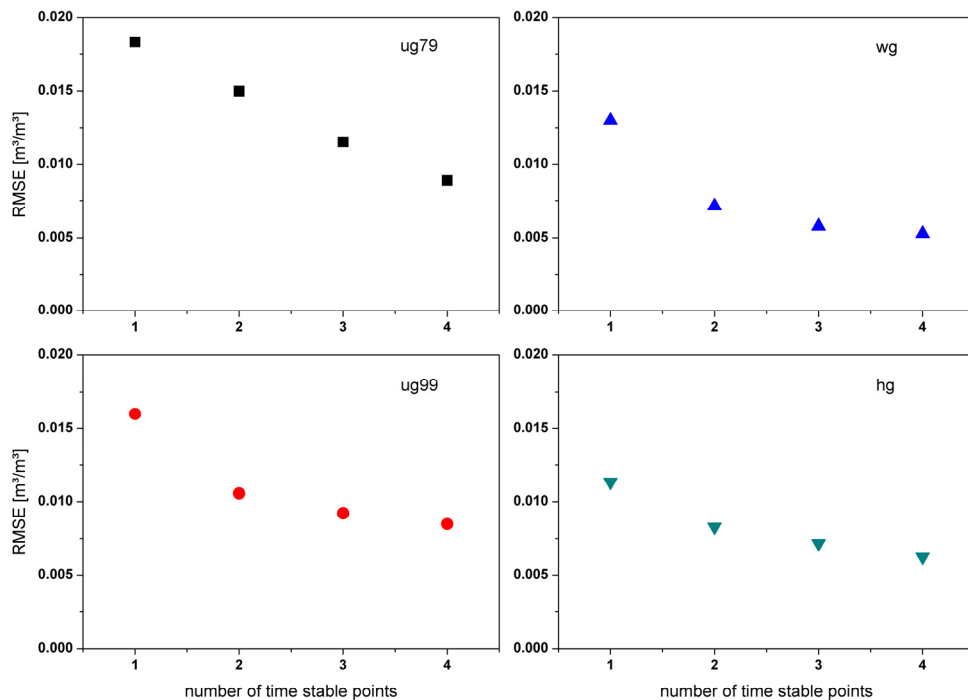


Figure 4.5. Effect of number of time-stable points on RMSE values.

The number of samples required to estimate mean soil moisture within 2% and 5% error of soil moisture is shown in Fig. 4.6. Only a few samples are needed when estimates of mean soil moisture with 5% accuracy are required. However, the number increases considerably when higher accuracy is required. On all sites, more samples are required under wet conditions.

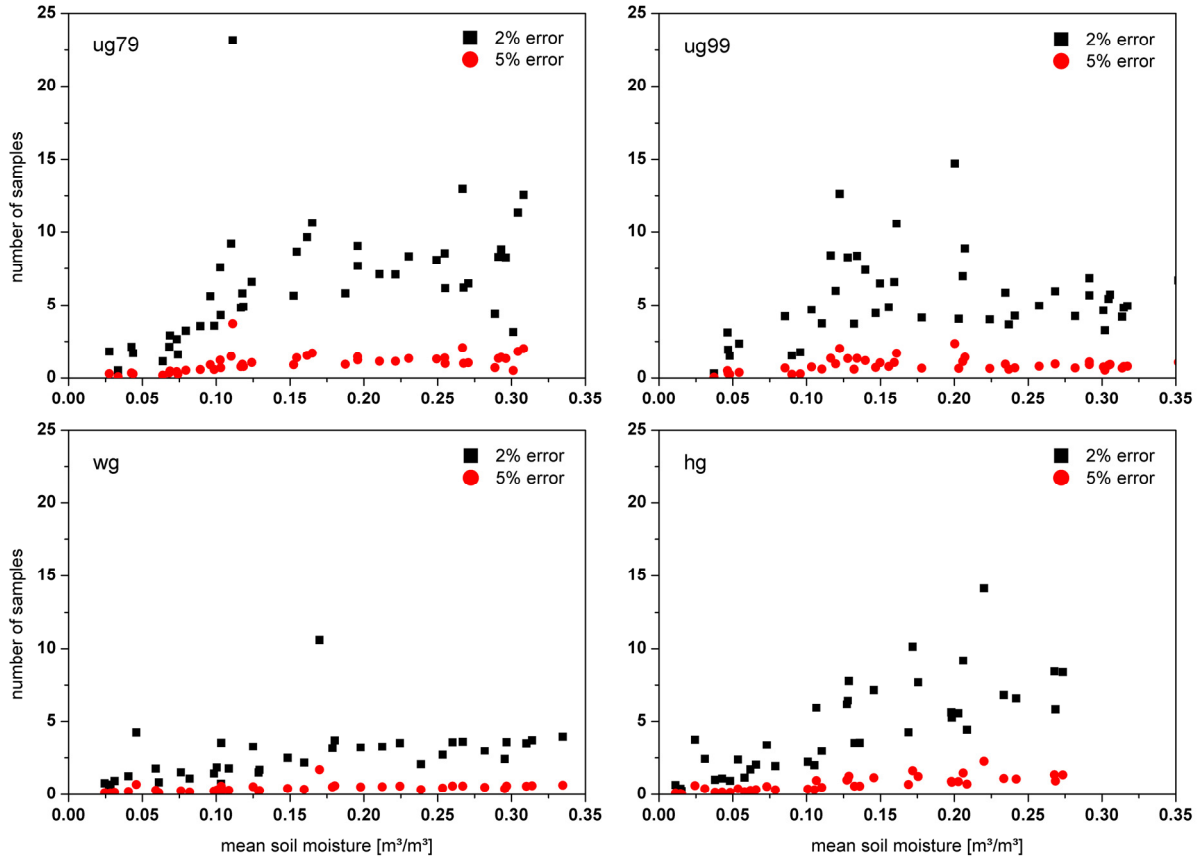


Figure 4.6. Number of time-stable sampling points required to estimate mean soil moisture with 2% and 5% accuracy.

The ranking of most time-stable points changed between 2004 and 2005 (Fig. 4.7). In some cases, the ranks shifted considerably, while other points remained close to their rank in 2004. Yet, while their rank changed, most points exhibited a similar behaviour with respect to mean relative difference and standard deviation. This is related to the fact that values of mean relative difference are very close (Fig. 4.2), which means that small changes of $\bar{\delta}_j$ cause considerable rank variation. This is in agreement with Martínez-Fernández and Ceballos (2003), who reported that the rank position of soil moisture monitoring sites may change between years. Nevertheless, the overall characteristics of the time-stable points, i.e. representing a wet, average or dry location, persisted for almost all sampling sites in their study.

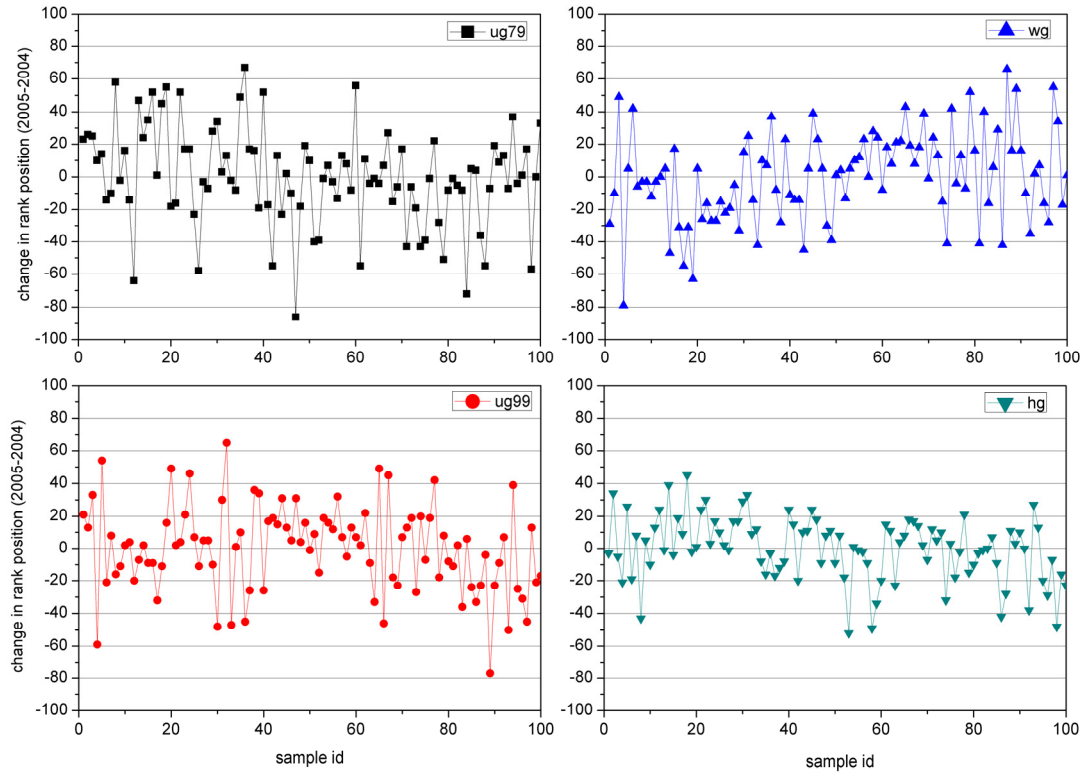


Figure 4.7. Change in rank position from 2004 to 2005.

Even if time-stability of individual points changes only slightly, the question arises if moisture patterns for entire sites are altered when single points change their rank position over time. The persistence of spatial patterns can be determined with a correlation analysis. The results of the Spearman rank correlation (Eq. 7) are presented in Table 4.5. The wg and ug99 site reveal higher values than the ug79 site, although the correlation is weak for both sites. Rank correlation on the ug79 and the ug99 site does not increase with recharge or discharge. Correlations on the wg site tend to rise after a few precipitation events in 2004. However, this effect disappeared in 2005. Compared to the other sites, correlation is persistently highest on the hg site. This indicates temporal persistence of moisture patterns between the sampling dates. As on the other sites, correlation did not depend on wetting or drying conditions. Compared to studies of Vachaud et al. (1985), Comegna and Basile (1994) and Gómez-Plaza et al. (2000), the correlation of ranks between the sampling dates is low. On the other hand, Mohanty and Skaggs (2001) found varying degrees of correlation depending on the surface characteristics of the study site, i.e. topography and vegetation cover.

The difference between temporal persistence at the point and field scale might be explained by surface properties. While individual time-stable locations occur on each of the sites, the field scale persistence is influenced by the variability of plant cover and composition. All sites are flat and with comparable exposition, so topographic effects on field scale persistence can be excluded (Schneider et al., 2008). Due to continued browsing by sheep, the hg site is very homogeneous regarding plant cover and plant composition. No major disturbances are apparent, so soil moisture

evolution is not influenced by heterogeneity of vegetation. On the other hand the wg, ug99 and ug79 site are more densely vegetated, and the vegetation pattern is more heterogeneous than on the hg site. Thus, soil moisture storage is influenced more strongly by vegetation. As plant growth changes the water demand during the vegetation period, also soil moisture patterns might change along with vegetation patterns. Single locations that are not subject to these changes might remain time-stable, but the temporal persistence of the spatial pattern might be affected. Gómez-Plaza et al. (2000) reported similar effects between vegetated and non-vegetated sample locations.

Table 4.5. Spearman rank correlation coefficients of consecutive soil moisture measurements in 2004 and 2005 and precipitation [mm] at IMGERS one day before soil moisture measurement.

Date	Precipitation [mm]	ug79	ug99	wg	hg
2004					
20.07.2004	0.0				
27.07.2004	1.6	0.088	0.191	0.103	0.186
03.08.2004	23.9	0.118	0.052	0.024	0.509
12.08.2004	60.7	0.097	0.457	0.211	0.655
17.08.2004	4.0	-0.019	0.495	0.074	0.672
23.08.2004	0.0	0.001	0.361	0.130	0.599
29.08.2004	7.1	-	0.268	0.039	0.566
02.09.2004	0.0	-	0.428	0.097	0.620
07.09.2004	0.2	0.258	0.287	0.089	0.665
14.09.2004	0.0	0.214	0.393	0.061	0.567
16.09.2004	21.2	0.127	0.178	0.222	0.441
21.09.2004	7.0	-0.004	0.089	0.264	0.442
2005					
08.06.2005	0.0				
16.06.2005	11.2	-0.029	0.145	-0.069	0.441
17.06.2005	0.0	0.156	0.170	0.152	0.718
18.06.2005	0.0	0.273	0.260	0.241	0.624
19.06.2005	0.0	0.274	0.346	0.229	0.578
20.06.2005	0.6	0.352	0.426	0.296	0.710
21.06.2005	0.0	0.466	0.355	0.303	0.705
22.06.2005	0.0	0.418	0.356	0.263	0.565
30.06.2005	15.4	0.038	0.112	0.203	0.519
23.07.2005	1.7	-0.003	0.259	0.198	-
27.07.2005	1.1	0.053	0.189	0.129	-
05.08.2005	0.0	0.163	-0.090	0.039	-

It is interesting to note that the sites with highest organic carbon content and a high clay fraction are not the most temporal persistent sites. Organic carbon and clay content on the hg site is lowest (Steffens et al. 2008), which reduces water holding capacity. However, soil moisture patterns are most persistent on this site. This might seem counter-intuitive but one has to consider that the atmospheric demand in such a semi-arid environment dominates water fluxes and hence the influence of soil properties is superimposed by high evapotranspiration rates. Note that the rank correlation increases when the time interval between the sampling days is short, as in June 2005. For shorter sampling intervals, temporal persistence is higher on the ug79, ug99 and wg sites. Also, the hg site reaches rank correlations higher than 0.7. Obviously soil moisture patterns do persist for a certain time throughout a discharge cycle, but these patterns do not remain when the time gap between the measurements becomes larger.

Identified time-stable points can be used to validate soil moisture measurements from satellite data. ERS soil moisture and ground measurements for 2005 and 2006 (Wagner et al., 1999) are compared in Fig. 4.8. The ERS data is presented in three ways: (1) The values for the closest pixel (ERS_c), i.e. where the center of the pixel is closest to the experimental sites, (2) the areal mean (ERS_{am}) and (3) the distance weighted (ERS_{wm}) soil moisture of the four pixels surrounding the experimental sites. ERS_{am} is the average of the four surrounding pixels while ERS_{wm} uses an inverse distance scheme considering the separation between pixel center and experimental areas.

The ground measurements comprise mean field soil moisture of the four grazing sites, and mean soil moisture derived from the four best time-stable points of each site. Mean values and standard deviations of the entire data set and of the time-stable points are comparable. In some cases, the standard deviation for the time-stable points is higher than for the complete data, but the effect is marginal. Therefore, the time-stable points provide a satisfactory estimate of the mean soil moisture conditions in the top layer of the four grazing sites. While these individual points seem valuable to predict soil moisture on a slightly larger (field) scale, the comparison between the mean moisture content from the time-stable points and remote sensing data is much less satisfying (Fig. 4.8). Obviously, bridging the gap to the much larger spatial scale of the scatterometer measurement introduces considerable bias. Correlations reach 0.6 at most and were highest between ERS_c and mean time-stable as well as mean field soil moisture and lowest between ERS_{am} and ground-based measurements (Table 4.6).

Table 4.6. Correlation matrix between satellite and ground-based soil moisture data.

	ERS_c	ERS_{am}	ERS_{wm}	field	ts
ERS_c	1.00				
ERS_{am}	0.97	1.00			
ERS_{wm}	1.00	0.98	1.00		
field	0.51	0.30	0.45	1.00	
ts	0.56	0.35	0.50	0.99	1.00

(ERS_c : ERS closest pixel, ERS_{am} : ERS areal mean, ERS_{wm} : ERS weighted mean, field: mean field soil moisture, ts, mean soil moisture from time stable points)

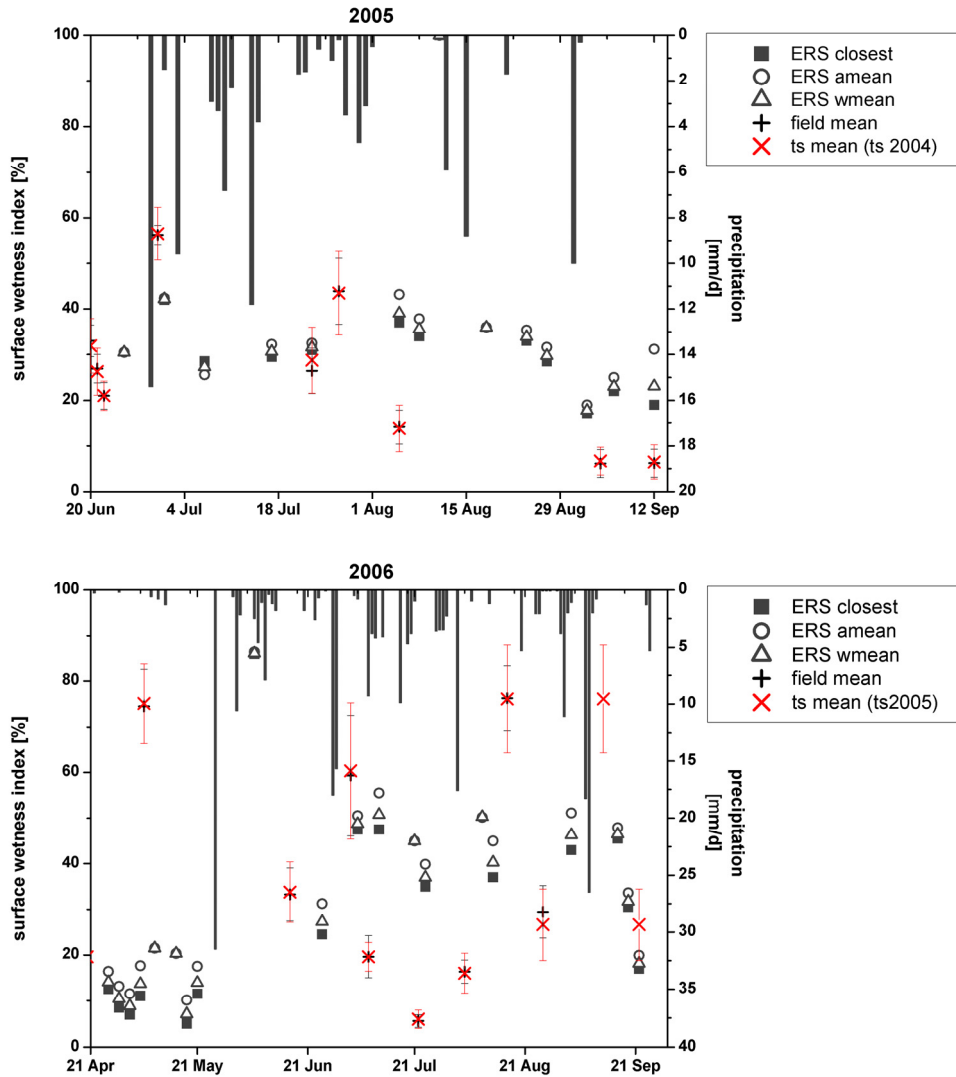


Figure 4.8. Comparison of ERS soil moisture and ground-based measurements in 2005 and 2006. ERS closest: single value of pixel closest to grazing sites; ERS a mean: mean values of four pixels adjacent to grazing sites; ERS w mean: inverse distance weighted mean values of four pixels adjacent to grazing sites; field mean: mean soil moisture of all sampling points on the four grazing sites; ts mean: mean soil moisture of selected time-stable points on the four grazing sites. Error bars indicate one standard deviation.

Unfortunately, ground measurements in a high spatial resolution at the footprint scale of the ERS sensor are not feasible. Considering the constraints for ground-based soil moisture measurements in the study region the sites selected for this study were chosen: (a) because land use, vegetation and topography are representative for the Xilin river catchment, (b) because the vicinity of the four sites allowed comparing the effect of different grazing intensities on soil moisture and temporal stability and (c) because the four sites represent the range of land management in the Xilin river catchment. By averaging information from these different sites the span of soil moisture on areas with different grazing intensities that affect the backscatter signal of the scatterometer can be captured. Other studies have presented satisfying agreements between satellite remote sensing and ground-

based soil moisture, even if ground data were sparse or did not cover the sensor footprint (Wagner et al., 1999; Pellarin et al., 2006). In our study, the ground and satellite measurements show similar soil moisture ranges on a few occasions, but on several other occasions the ground measurements show much lower or higher soil moisture. At this stage, only a few concurrent ground and satellite measurements are available, so it is not possible to evaluate the comparability of the two measurement scales in detail. As the study site is a grassland area, large reflectance errors due to surface inhomogeneities are not likely. Therefore, we assume that rainfall variability within the satellite footprint is a driving factor for the observed differences. Precipitation during the vegetation period is mainly of the convective type. Thus, locations only a few kilometres apart can receive different amounts of rainfall during a storm. In combination with the scale gap between the measurements, the rainfall patterns in the study region might lead to the partly huge bias between ground and satellite soil moisture values. This scale problem might be dampened when downscaled radar data is available (e.g. from the advanced scatterometer on board the MetOp satellite).

4.4 Conclusions

The analysis of soil moisture data sets with respect to time-stability resulted in the identification of representative sample locations to predict average field soil moisture. With only a few points selected from a total of 100 sample points, mean soil moisture could be accurately estimated. The results show that the measurement effort in long term studies can be reduced when time-stable locations occur. Despite this successful identification, the concern remains which effort and temporal duration is necessary to derive reliable time-stable sample points. From the data presented in this study, we conclude that one year of soil moisture data was sufficient to derive reliable time-stable points. The combination of two years of soil moisture data did not result in an improvement in the independent third year.

Time-stability on the field scale proved to differ between the four sites: while the hg site showed high temporal persistence of spatial patterns, the persistence on the other sites was only weak. We relate these differences to surface properties of the four sites. Although no detailed analysis of vegetation composition was performed, the hg site is clearly the one with little vegetation and very uniform vegetation composition and patterns. This homogeneity is reflected in the rank correlation analysis.

The application of time-stable points to validate large scale remote sensing products in the Xilin river basin is not feasible at this stage. The preliminary results give a two-edged picture. Time-stable points could estimate radar soil moisture satisfactorily in some cases, but failed completely in other cases. The different measurement scales seem to promote bias introduced by rainfall variability. Further comparison has to prove whether the time-stability method carries the potential to validate satellite soil moisture information in the Xilin river area. The resolution of the scatterometer footprint will be higher in the future, so the scale gap between ground and satellite measurements might be easier to overcome.

This paper attempted to provide further insight on the potentials and limitations of the time-stability concept. Only on the hg site soil properties such as sand content or

organic carbon content explained temporal persistence of soil moisture sufficiently. The energy balance and the resulting atmospheric water deficit during the vegetation period levels the differences that might occur due to vegetation and soil characteristics on the four sites. The results showed that a careful application is necessary, especially when accurate information is required. Also, the validity of the concept when the spatial scale of atmospheric processes is smaller than the sampling scale needs to be determined. Nevertheless, the methodology proved to facilitate the monitoring of wetting and drying cycles.

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Erklärung

Ich erkläre: Ich habe die vorgelegte Dissertation selbständig und ohne unerlaubte fremde Hilfe und nur mit den Hilfen angefertigt, die ich in der Dissertation angegeben habe.

Alle Textstellen, die wörtlich oder sinngemäß aus veröffentlichten Schriften entnommen sind, und alle Angaben, die auf mündlichen Auskünften beruhen, sind als solche kenntlich gemacht.

Bei den von mir durchgeführten und in der Dissertation erwähnten Untersuchungen habe ich die Grundsätze guter wissenschaftlicher Praxis, wie sie in der „Satzung der Justus-Liebig-Universität Gießen zur Sicherung guter wissenschaftlicher Praxis“ niedergelegt sind, eingehalten.

Gießen, im September 2008