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Soil Moisture Deficits in South-Central Sweden

I – Seasonal and Regional Distributions

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Some commonly used assumptions about climatically induced soil moisture fluxes within years and between different parts of a region were challenged with the help of a conceptual soil moisture model. The model was optimised against neutron probe measurements from forest and grassland sites. Five 10 yrs and one 105 yrs long climatic records, from the province of Östergötland, situated in south-central Sweden, were used as driving variables. It was concluded that some of the tested assumptions should not be taken for granted. Among these were the beliefs that interannual variations of soil moisture contents can be neglected in the beginning of the hydrological year and that soils usually are filled up to field capacity after the autumn recharge. The calculated climatic induced dryness was estimated to be rather insensitive to the choice of climatic stations within the region. Monthly ranges of soil moisture deficits (1883-1987) were shown to be skewed and it is therefore recommended to use medians and standard deviations in statistical analyses of "normal" ranges of soil moisture deficits.

Introduction

The soil moisture dynamics is a key to the water balance of a region. The hydrological conditions of the soil moisture zone determine if rainfall and snowmelt will lead to groundwater recharge and to what degree the atmospheric evapotranspiration demands can be satisfied.

In spite of this, soil moisture is often handled with crudity in water balance

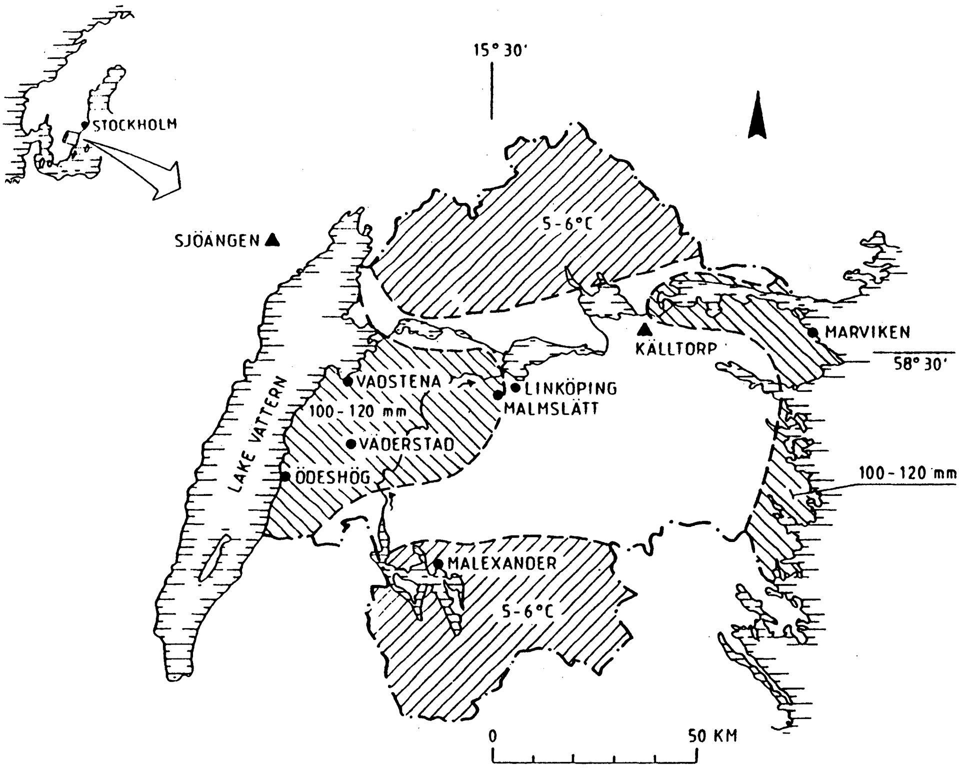

Fig. 1. Östergötland. Location of climatological stations (●) and soil moisture measurement sites (▲). [\\\] = areas with average precipitation April-June below 120 mm (otherwise 120- 140). [///] = areas with 30-year average temperature below + 6°C (otherwise + 6°-+ 7°C). Average values were calculated, using data from 1931-1960.

studies. Measurements of soil moisture are costly and time consuming. The spatial heterogeneity is considerable both in micro, meso and macro scales. Because of these difficulties, many hydrologists lack the tradition to include a serious analysis of soil moisture dynamics in the analysis of basin behaviour.

During the latest decade, physically based soil moisture models, which rely on a large number of parameters and driving variables have been developed (*e.g.* Jansson and Halldin 1980). However, those models are limited to applications where the model requirements can be fulfilled. For many hydrological purposes, the need of information about soil moisture dynamics would be met by simpler models. Therefore, it is necessary to complement the physically based models with simpler conceptual soil moisture models, with limited input requirements. To be useful, however, such models must include descriptions of physical and biological processes that, in spite of the limited input data requirements, are acceptably close to reality.

In this study, such a model has been used to testify some common assumptions of climatically induced soil moisture dynamics, including the importance of regional variability of the climatic pattern, the timing of dryness culmination, dryness and the hydrological new year, the influence of snowmelt on soil moisture dynamics, the recharge up to field capacity during autumn, and the characteristics of the statistical distributions of soil moisture deficits. The analysis was made for the province of Östergötland, situated in south-central Sweden (Fig. 1). The main regions are the central plains, which have a large fraction of clayey soils, and the areas north and south of the plains, which are dominated by forest and mainly situated on till soils. Östergötland is among the driest parts of Sweden, with an annual average (1931-1961) precipitation of 630 mm and actual evapotranspiration of 450 mm (Falkenmark 1976).

Method and Data

Simulation Model Used

Soil moisture deficits in the upper metre of the soil profile were calculated from climatic data by a mathematical and operational model, based on some well known concepts. The model was optimised against neutron probe soil moisture data. The used equations are fully described by Andersson (1988). The model has been constructed with the aim to get a tool which is possible to use for time periods and sites where the only available data are information of precipitation and air temperatures plus monthly means of potential evapotranspiration.

Routines for percolation during days with rainfall or snowmelt, actual transpiration, snow accumulation and melting (degree-day equation) were taken from the runoff model HBV (Bergström 1976). Contrary to drainage routines in the HBV model, soil moisture contents were allowed to exceed field capacity in the soil moisture model, and the parameter *SMAX* which is described as field capacity in the HBV model, represents the water content when the soil is fully saturated;

$$\frac{\mathrm{PERC1}}{\mathrm{INSOIL}} = \left(\frac{SM}{SMAX}\right)^{\beta} \tag{1}$$

where

PERC1 – percolation (mm)
INSOIL – infiltrating rainwater or snowmelt (mm)
SM – calculated soil moisture storage above wilting point (mm)
SMAX – the water content above wilting point at full saturation (mm)
β – parameter

In general, the parameter β was set so that the percolation was close to zero at field capacity. Field capacity was defined as the typical measured moisture content in the upper metre of the soil during the time between the end of the autumn recharge and the start of the period with significant transpiration demand. Since many soil moisture measurements were available from this period, this approach was chosen instead of using field capacity estimates from the pF curves. The water content at soil saturation (pF 0.7) and wilting point (pF 4.2) were, however, estimated from pF-curves, evaluated from laboratory soil pressure potential measurements made at soil samples from the various sites (Andersson and Harding in manuscript). From Eq. (1), percolation is only calculated during days when rainfall or snowmelt occurs. During days when soil moisture exceeds field capacity, drainage is calculated as the sum of Eqs. (1) and (2)

$$\mathrm{PERC2} = \mathrm{MAXPERC}\ \frac{SM-FC}{S\mathrm{MAX}-FC} \tag{2}$$

where

PERC2 – percolation (mm)
*MAX*PERC – parameter, representing daily percolation at soil saturation
FC – field capacity (mm).

Experiences from soil moisture simulations for the Velen forest (Andersson 1988; Andersson and Harding in manuskript) showed that significant improvements were achieved if unsaturated macropore flow was considered. Otherwise, soil moisture was clearly overestimated after rainstorms in the autumn recharge period. Therefore, a routine for macropore flow was incorporated into the soil moisture model. The routine is a simplified accounting procedure where recent rainfall conditions determines if macropore flow is initiated, and the accumulated sum of soil moisture deficits determines the present degree of macroporosity. Macropore flow is calculated as a function of rainfall, antecedent conditions and a site dependent parameter, optimised against soil moisture data from recharge periods.

Actual daily evapotranspiration in mm/day (*Ea*) was considered to be equal to the monthly long time average potential evapotranspiration for each month (Ep_m) until a certain soil moisture deficit (% of *S*MAX), determined by the parameter *Lp*, existed and the ratio between actual and potential evapotranspiration became determined by a root constant function;

$$\text{when } SM > Lp\ S\mathrm{MAX}:\ Ea = EP_m \left(\frac{SM}{Lp \times S\mathrm{MAX}}\right) \tag{3}$$

The parameter *Lp* was estimated by optimization. *Lp* was found to be high for short rooted vegetation where the water stress is increased as soon as the upper parts of the soil profile are depleted, and low for deep rooted trees which can maintain sufficient water supply from deeper soil layers after the upper soils layers have dried out.

When the model was tested against measurements from a grass covered clayey soil, it was evident that transpiration rates during summer were increased after rainstorms, that occurred during periods with high moisture deficits, although the total water content of the soil still was low. This could not be explained by macropore flow, since the model managed to predict the moisture recharge, but underestimated the later recession. Therefore, when used for the grass covered clayey soil, the linear root constant was complemented with a more dynamic layer function. An "upper transpiration zone" (*UTZ*) was fed by rainfall in times when the moisture content was below the threshold value, at which actual evapotranspiration otherwise is limited by the integrated water content over the upper meter (*Lp*). Due to the exponential decrease of hydraulic conductivity that occurs when the soil dries out, the maximum content of the upper transpiration zone for day *i* (mm) was determined by the difference between the actual water content and the water content at *Lp*. The zone is emptied by the calculated transpiration and by a downward flow, set to 10% of the water content in the zone;

$$\text{when } SM_{i-1} < Lp: \quad UTZ_i = UTZ_{i-1} + P_i - Ea_{i-1} - (0.1 \times UTZ_{i-1}) \qquad (4)$$

$$UTZ_i \text{ is maximized to } Lp - SM_{i-1}$$

The assumption of the same potential transpiration, independent of years might be questioned. However, studies (Calder *et al.* 1983; Andersson and Harding in manuscript) have shown that the use of climatological mean estimates of potential evapotranspiration often gives better agreements between simulated and measured soil moisture deficits than achieved with daily estimates using the Penman equation (Penman 1948). The paradox that more detailed calculations of the atmospheric moisture demand, lead to less accurate predictions of soil moisture deficits was probably due to the neglecting of biological feedbacks. The Penman equation is entirely based on weather variables and takes no account of strong negative feedback response that exists between stomatal resistance and atmospheric humidity during dry spells (*e.g.* Roberts 1983). However, the use of more sophisticated evaporation equations, has not improved soil moisture simulations. This may be due to difficulties to make accurate determinations of their empirical constants (Calder *et al.* 1983). The rather conservative nature of transpiration is thus well described by the use of average potential values. It can be argued that the Penman equation is unsuitable for forest conditions. However, it has been shown that the Penman equation, when used in conjunction with an optimised equation that calculates the ratio of actual transpiration to the Penman potential estimate, also has value in estimating transpiration losses from forest (see *e.g.* Calder and Newson 1979; Calder 1982; Calder 1985). Interception losses are, however, not well described by the Penman approach, and have to be considered separatly (see *e.g.* Shuttleworth and Calder 1979).

Forest interception was simulated with an exponential relationship, involving daily rainfall and considering the fraction of the day when canopy is wet (and no transpiration is considered to take place). The parameter values of mean rainfall intensity and of the ratio between the time that the canopy is wet during, and immediately following, rainfall and the number of rain hours were estimated as the average for U.K. conditions (Calder and Newson 1979), which probably does not differ very much from conditions in south-central Sweden. The interception equation was optimised against interception measurements from the Velen forest (Bringfelt and Hårsmar 1974), to give appropriate annual interception ratios. Comparison with interception measurements showed that the model was very well adapted to simulate interception (Andersson 1988). Interception was only calculated for the forest sites.

A temperature effect routine simulated the reduction of actual transpiration rates during time periods in spring and early summer when soil and air temperatures can act as limiting factors on the development of the vegetation and on the evapotranspiration. The index is a linear function of the number of days with a five days average air temperature above +5° C. The number of days until the temperature ceases to affect the transpiration rates was held at a value of 28 (four weeks). After that, the index has reached the value of 1.0, which means that if not hindered by water stress, transpiration can occur at the meteorologically determined potential rate.

Optimization and Validation of the Model

Neutron probe soil moisture data, representing the integrated moisture volume fraction over the upper metre of the soil profile, from sites which were considered to be representative for the main soil and vegetation regions of Östergötland, were used to optimize and validate the model.

Neutron probe measurements from Källtorp (Milanov 1982) (Fig. 1) were considered typical for the plains. The site consisted of grassland on clayey loam in the upper half and clay in the lower half of the top metre. After autumn recharge, the groundwater level was on average one metre below the soil surface. The average of the readings from three tubes was used for the model optimization. Time intervals between the readings varied between a few days and a month. R^2 values of 0.92 for the optimization (1976-1978) and 0.77 for the validation period (1979-1981) were obtained.

Readings from two forest and one grassland tubes at Sjöängen (Velen's representative basin) (Forsman and Milanov 1971) (Fig. 1) were considered typical for the forest and till soil dominated areas. Soil (sandy loam) and topographic conditions did not significantly differ between the three tubes. Groundwater tables were relatively shallow (on average one metre below soil surface after the autumn recharge). The time interval between the readings varied between a week and a month. Separate parameter optimizations were made against the readings from the three sites. R^2 values varied between 0.86 and 0.94 for the optimization period (1968-1970). A validation period (1971-1973) was only available for the two forest tubes (R^2 values of 0.83 and 0.85). Applications of the model at Velen are further

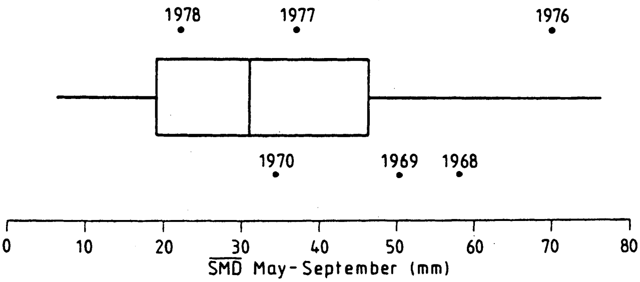

Fig. 2. Box-plot showing the median, quartiles, maximum and minimum values of soil moisture deficits May-September (1883-1987), calculated as the mean from the two forest and the two grassland optimizations. The position of the years used for optimization are shown above (Källtorp), and below (Sjöängen) the box-plot.

discussed by Andersson (1988).

It is important that data used for model optimization covers as many different types of combinations of climatic situations as possible. The years that were used for model optimization were included in a generally dry period (Fig. 2). In spite of that, the used years had a rather good distribution between wet and dry, although none of the years used for optimization at the Velen forest were wetter than the median (1883-1987). However, no systematic over- or under-estimations of deficits were found during the two wet summers of 1971 and 1972, used for model validation. This indicates that the model, with the chosen sets of parameter values, was not biased towards dry conditions.

Driving Variables

Driving variables for model optimization and validation, consisted of daily precipitation and temperature data from stations nearby the soil moisture tubes, plus estimates of monthly means values of daily potential evapotranspiration (1931-1960) from Linköping for the Källtorp optimization and from Örebro for the Velen optimization (Wallén 1966).

A 105 yrs climatic record from Linköping (1883-1987) and five 10 yrs records (1974-1983) from the region were used (Fig. 1). When using a 105 yrs record of precipitation data, it is probable that some degree of inhomogeneity has to be faced. In order to construct a homogeneous record, correction factors were calculated by a method which made it possible to analyze if there had been a significant change in the relative distribution of precipitation events in classes of various rainfall amounts. The method is described by Andersson (1989).

Analysis of Some Generally Used Assumptions

Some assumptions that often are used in water balance studies, were challenged with the help of the described soil moisture model and climatic records.

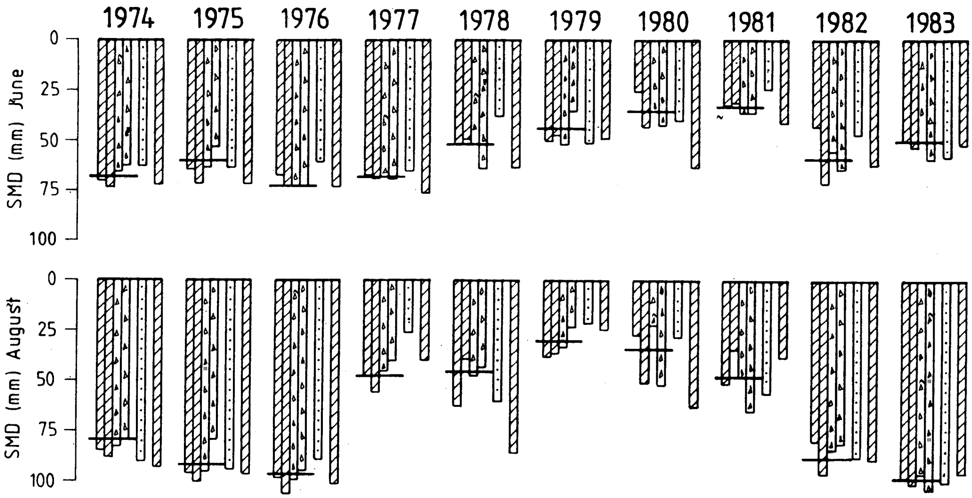

Fig. 3. Calculated maximum soil moisture deficits during June and August (1974-1983), using climatic data from different stations. From left to right; Ödeshög, Vadstena, Väderstad, Malmslätt, Malexander and Marviken. The calculated deficits when the mean precipitation from four stations was used is represented by a line through these four columns. ▨ = drier areas, close to lake Vättern or the Baltic. ⁙ = the wetter highlands. ◿ the central areas.

Regional Variability of the Climatic Pattern

Among hydrologists, there exists a consciousness of the importance of the spatial variability of precipitation when calculating runoff. It could therefore easily be suggested that climatic spatial variability would be equally important for regional soil moisture variability.

Different parts of Östergötland have somewhat different climatic characteristics (Fig. 1). The entire area, but especially the eastern part is characterised by dry periods in the early summer. The yearly precipitation in this region and in the surroundings of lake Vättern is lower than the other parts of the province. Higher situated areas in the north and south have on average lower air temperatures than the rest of the region. The area close to the Baltic is characterised by long and mild autumns.

In order to examine the influence of regional climatic variability, ten years of simulations (1974-1983) were compared, using the Källtorp optimization and climatic data from six stations in the province of Östergötland (Fig. 1). It was shown that regional climatic variability led to some spatial variations of the maximum soil moisture deficits (Fig. 3). However, the pattern was not static. Climatic stations that one average had the lowest summer precipitation, and thus generally generated the highest calculated soil moisture deficits, could be among the wettest for single years (Fig. 3).

Contrary to other hydrological extremes, drought is the result of extreme non events (Yevjevich *et al.* 1983). This means that the probability that a regional diversification will develop is small during dry spells as a consequence of the few occasions when the diversification can be enlarged. This was reflected in the fact that the standard deviations between maximum deficits, calculated with precipitation records from different stations, were larger during wet years (8.8 mm in June, 12.6 mm in August) than during dry years (5.0 mm in June, 5.3 mm in August). Regional climatically induced variations are, however, reduced not only in times of prolonged dryness, but also after prolonged wet periods when the soil is close to saturation. For example, the climatically induced spatial soil moisture variations increased from an insignificant level in early spring (*i.e.* when most soils are close to saturation) to a maximum in August.

Often, both for hydrological and agricultural purposes, the interest is focused towards areal estimates of soil moisture deficits. One could suspect that using rainfall data from only one station would give too skewed estimates to be useful for regional calculations. However, calculations of monthly maximum deficits, using means of precipitation data from four stations situated on the plains, did on average not differ with more than 4.5 mm in June and 6.5 mm in August from calculations, for each single station (Fig. 3).

Dryness Culmination

It is often taken for granted that the soil is driest during the middle of the summer. This is the period when the shallow soil layers have dried out and the effects of water deficits on shallow rooted vegetation are visible.

The yearly dryness culmination in the soil moisture zone was calculated, using one grassland and one forest optimization from Sjöängen and the 105 yrs climatic record. It was shown that yearly maxima of soil moisture deficits could occur any time between the middle of June and the end of October (Fig. 4). Maximum deficits occurred later in the forest soil than in the grassland soil because water outputs continued to be larger than inputs for a longer time in the forest. This was caused by high transpiration being maintained even in times when the upper soil layers were depleted of water. In addition, interception losses and the possibility of macropore flow from an unsaturated soil profile were larger (Andersson 1988).

Dryness and the Hydrological New Year

When calculating water balances, it is advantageous if the storage terms can be assumed to be constant from year to year. Therefore, water balances are usually made for hydrological years, starting when interannual differences of the storage component are believed to be as small as possible. In Sweden, the 1st of September or the 1st of October are used as the hydrological new year (see *e.g.* Waldenström 1976).

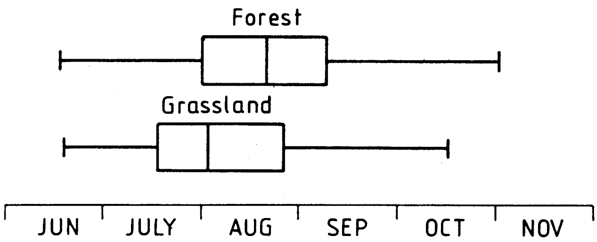

Fig. 4. Time periods for the calculated (1883-1987) yearly maximum of soil moisture deficits using model optimization from the driest forest site and the grassland site at Sjöängen. The box-plot graph shows the median, quartiles and the extent of the distribution.

Unfortunately, the interannual variations of soil moisture deficits (1883-1987) were never as high as during this time of the year, with an interquartile range of 40 mm, calculated as the mean of the simulated deficits from the four used model optimization (Fig. 5). The range was largest for the forest where larger deficits could develop. In the beginning of the hydrological year, interannual soil moisture deficits had a range from moisture contents that exceeded field capacity to a maximum deficit of 125 mm for the grassland soil and up to 180 mm for the driest forest soil (Fig. 6). It is thus possible that considerable errors will occur if a water balance is calculated on a yearly basis, using the early autumn as the hydrological

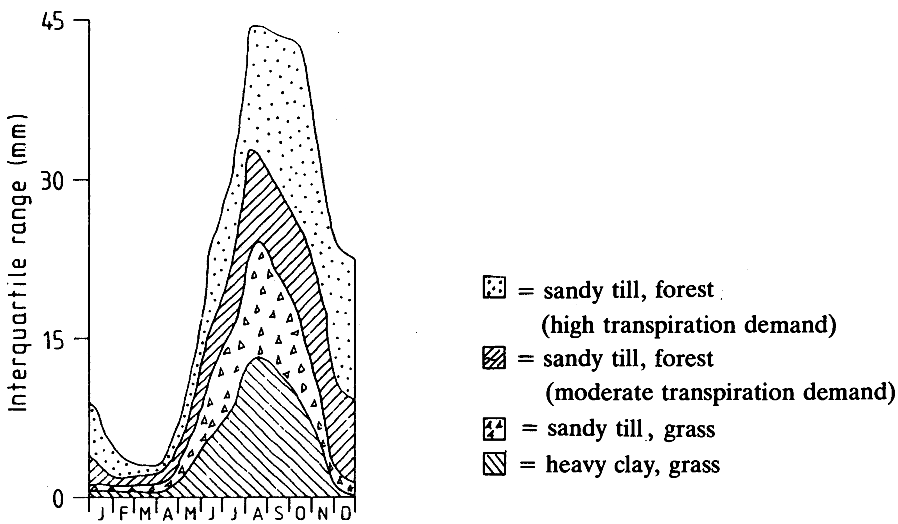

Fig. 5. The average (from two forest and two grassland model optimizations) interquartile range of monthly means of soil moisture deficits (1883-1986). The contribution to the mean interannual variations from the four different soil and vegetation types during different parts of the year are shown.

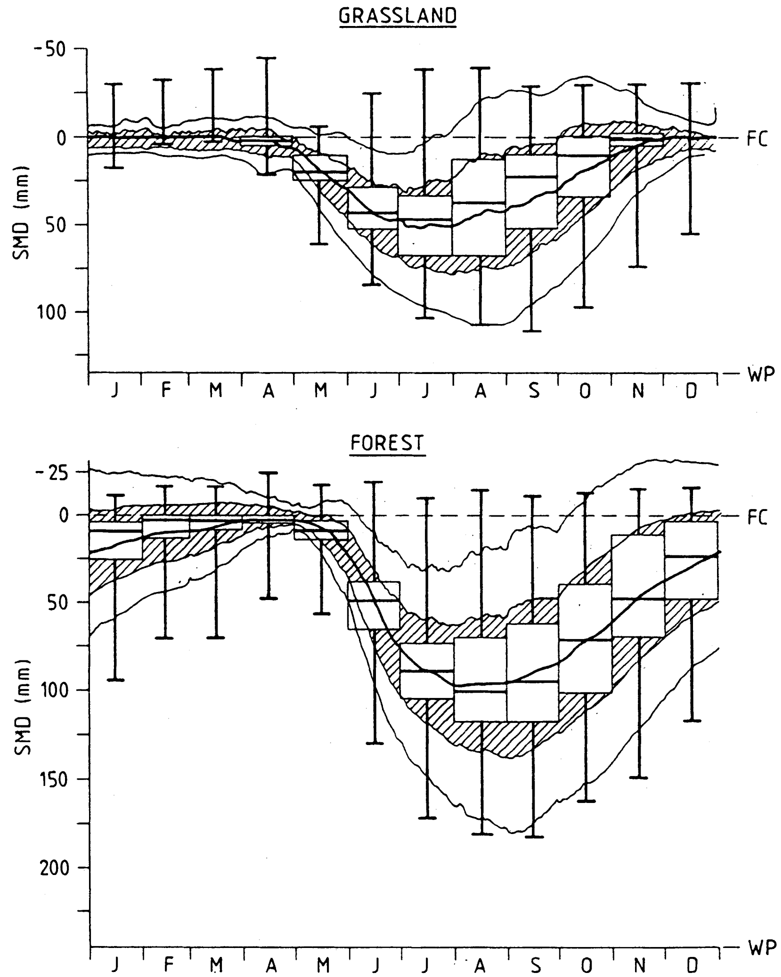

Fig. 6. Calculated daily means +/− one and two standard deviations of soil moisture deficits (1883-1986), using optimizations from grassland on clayey soil (Källtorp) and from the driest of the forest sites on till soil (Sjöängen). The boxes represents monthly medians and quartiles. The range between the daily maximums and minimums for each month are also shown.

new year, and considering the soil moisture term as constant from year to year.

Statistical Analysis

In a statistical analysis of the normal range of the distribution of events, average values and standard deviations are the most commonly used parameters. However, average and standard deviations should only be used when it can be shown that the studied events are normally distributed.

Averages and standard deviations for soil moisture deficits (1883-1987) were compared to median and quartiles. It was shown that an unsymmetrical distribution of monthly deficits existed (Fig. 6). When the forest optimizations were used, a few dry years caused the mean to be significantly lower than the median during autumn and spring. For all sites, except the driest forest site, some years with dry summers caused a summer mean that was lower than the median. For the driest forest site, however, a few unusually wet summers made the mean deficit smaller than the median. Thus, the use of means and standard deviations does not give good estimate of the distribution of typical values (Fig. 6). The skewed distributions make it preferable to use medians and quartiles when describing the normal range of distribution of soil moisture deficits.

The Influence of Snowmelt on Soil Moisture

It is often assumed that soil moisture conditions in spring, which includes the influence of snowmelt, are of significant importance for soil moisture availability in early summer. To test this, a regression between mean soil moisture deficits in spring (April, respectively May) and mean soil moisture deficits in June was made, using the four earlier described model optimizations. No significant correlations were found. The highest correlation ($R^2 = 0.38$) was found between mean deficits of May and June, using the optimization from the driest of the forest sites.

Saturation up to Field Capacity

The general view of yearly soil moisture fluctuations gives a picture with deficits during the summer months and recharge up to field capacity during autumn. However, the simulations showed that it is not uncommon with exceptions from this. Saturation up to field capacity could occur in all months, even at the driest of the forest tubes (Fig. 6). The assumption that the soil is filled up to field capacity towards the end of the year was shown to be generally true for the grass covered soils. It was only for a few years that a soil moisture deficit existed in the top metre of the soil at the end of December (1886, 1914, 1983) and even for those years, the deficits were small. However, according to the analysis, in the forest considerable deficits often existed towards the end of the year (Table 1).

Table 1 – Percentage of years (1883-1986) when a soil moisture deficit (mm) existed at the turn of a year according to time series optimised against soil moisture measurements from Sjöängen (sandy till) and Källtorp (clay).

	Deficit (mm)	1-10	11-50	>50
Forest:	Sjöängen 8	18	25	7
	Sjöängen 7	5	35	13
Grassland:	Sjöängen 6	1	0	0
	Källtorp	2	1	0

Conclusions

Many soil moisture dynamic related assumptions that often are used by routine should probably be handled with more care. Among these are the beliefs that interannual variations of soil moisture contents can be neglected in the beginning of the hydrological year (Fig. 5), and that soils follow a more or less static pattern of deficits during the summer and a filling up to field capacity after the autumn recharge. The latter assumption is especially questionable for forest soils where high soil moisture deficits can be developed even at deeper soil horizons and where recharge is slow because of interception and the influence of macropore flow. Dryness culmination was shown to occur at any time between the middle of June and the end of October with a median in early August for grassland and late August for forest (Fig. 4). It was shown that even for the driest of the forest sites, field capacity could be reached during any month (Fig. 6). The soil moisture conditions during spring, which includes the influence of snowmelt, were shown to be of minor importance for soil moisture conditions in early summer. It was also demonstrated that time series of soil moisture deficits are often skewed (Fig. 6), from which follows that variability should be described by medians and quartiles and not by means and standard deviations.

The errors introduced when using point precipitation values instead of areal mean estimates when calculating monthly maximums of soil moisture deficits in the region were small (Fig. 3). However, regional variations are also determined by other factors, like topography, soil characteristics and land use, which produces a landscape mosaic where every single part has its own way of responding to rainfall or snowmelt events. It must also be taken into consideration that the regional study only described the situation during a period of ten specified years in a region with rather small overall differences.

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