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Interception of a Dense Spruce Forest, Performance of a Simplified Canopy Water Balance Model

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The process of interception was studied in 25-year-old dense stands of Norway spruce in South Sweden. The throughfall was measured intensively during one month and extensively during four growing seasons using water captured by large roofs and with randomly distributed funnel gauges. It was found that about 45% of the precipitation was lost as interception loss from this dense forest canopy. However, many sources of potential error, particularly in measurement of precipitation and throughfall, may be involved in quantifying the interception loss. The data set was used to test the interception part of a hydrological model, SOIL. The model uses a simple threshold formulation to calculate the accumulation of intercepted water in a single storage variable. The model was able to estimate fairly well the long-term cumulative interception loss from the forest canopy. However, similarly to many other models, SOIL showed a pattern of overestimation of the interception loss during events with small precipitation and underestimation during events with large precipitation. It was concluded that the storage capacity was of major importance in modelling of long-term interception loss. Tree canopy water storage capacity on a leaf area basis was estimated to 0.7 mm which was three times larger than that obtained from a precipitation/throughfall graph.

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Introduction

The increased proportion of highly productive coniferous forest in Sweden as well as in many other countries during last decades may alter the water balance. During periods of precipitation, a quantity of moisture is caught and stored on the vegetation by the process of interception. Loss of precipitation intercepted in forest canopies can be as large as 50% of the precipitation on mature temperate coniferous forest (Anderson and Pyatt 1986). One can speculate that a larger leaf area in a denser stand leads to a larger interceptive and transpirative surface, which in turn leads to drier soil moisture conditions. Similarly to many others, Alavi and Jansson (1995) found "drier" soil moisture conditions for stands with higher aboveground biomass. This means that the drought sensitivity may increase as stand grows.

Parameters governing the interception process are not only stand parameters but also the distribution and duration of precipitation. Thus estimation of interception losses requires detailed measurement of precipitation and simultaneous observations of other variables likely to influence the loss process. The introduction of computer modelling during the 1960s and 70s was a big step forward on understanding the physical processes of interception loss. A significant contribution was the "Rutter" model (Rutter and Morton 1977; Rutter *et al.* 1971, 1975). This model regards the canopy as a single storage container and calculates the running water balance. However, the Rutter model requires hourly recording of rainfall and three-hourly recording of the meteorological variables governing evaporation, data which are not available at many sites. In order to achieve better practicability, Gash (1979) developed a semi-empirical simplification of the Rutter model by replacing its numerical approach with an analytical one. This model uses daily records of meteorological variables as input data.

In most hydrological models, as for example the SOIL model by Jansson and Halldin (1979), submodels for interception are included. In the SOIL model, a simple threshold formulation is used to calculate the accumulation of intercepted water using a single storage variable. The model can use both daily and within day input data. The SOIL model has been applied to a wide variety of vegetation types, soils and climatic conditions (*cf.* Jansson and Gustafson 1987; Johansson and Jansson 1991; Lewan 1993; Persson and Lindroth 1994; Alavi and Jansson 1995; Gärdenäs and Jansson 1995). However, the interception part of the model has never been explicitly tested on data with high temporal resolution. Up to now, the only test of this part of SOIL is one performed by Persson and Lindroth (1994). They tested it against monthly throughfall in a short-rotation willow stand and found that the model generally overestimated the interception loss.

The aim of this paper is to quantify the amount of interception loss from a dense spruce forest and calibrate and test the interception part of the SOIL model. Measured values of throughfall in different plots and periods were used.

Materials and Methods

Theory

Interception loss is defined as precipitation that is intercepted by trees and evaporated before reaching the forest floor. A simple water balance expression can be used to define the terms involved in the interception process on a specified time scale

$$I = P - T \quad (1)$$

where I is interception (mm), P is precipitation (mm) and T is the sum of throughfall and stemflow (mm). Throughfall has two possible paths, either direct throughfall: when the precipitation falls through the canopy without striking the vegetation, or indirect: when precipitation retained in the canopy later drains to the forest floor as canopy drip or stemflow.

The SOIL model assumes the whole tree canopy as a single storage unit where all precipitation will be temporarily stored before it drains to the forest floor or evaporates. Therefore, all throughfall occurs as indirect throughfall in the SOIL model. The interception water balance is modelled using a threshold formulation as shown below

$$S_{max} = i_{LAI} LAI \quad (2)$$

where S_{max} is the saturated storage capacity, i_{LAI} is a parameter and LAI is the leaf area index. Precipitation intercepted in each time interval, S_{int}, is either S_{max} minus the interception storage remaining from the previous time step, $S(t\text{-}1)$, or the precipitation, P, whichever is the lesser

$$S_{int} = \min\left\{P, \frac{(S_{max} - S(t-1))}{\Delta t}\right\} \quad (3)$$

Throughfall is then

$$T = \max\{0, P - S_{int}\} \quad (4)$$

The potential evaporation rate, E_p, from intercepted precipitation was calculated from the Penman combination equation in the form given by Monteith (1965) but modified for the wet canopy condition through excluding the canopy resistance. The canopy resistance is likely to be zero under wet canopy conditions (Stewart 1977; Teklehaimanot and Jarvis 1991).

$$\lambda E_p = \frac{\Delta R_n + \rho_a c_p((e_s - e)/r_a))}{\Delta + \gamma} \quad (5)$$

where R_n is net radiation, e_s is the vapour pressure at saturation, e is the actual vapour pressure, ρ_a is air density, c_p is the specific heat of air at constant pressure, λ

is the latent heat of vaporisation, Δ is the slope of saturated vapour pressure *versus* temperature curve, γ is the psychrometric constant and r_a is the aerodynamic resistance to the transport of water vapour from the single source point to a reference height. The reference height was assumed to be at 1 m above the canopy. The aerodynamic resistance is calculated using the equation given by Monteith and Unsworth (1990)

$$r_a = \frac{\left(\ln((z_{\mathrm{ref}} - d)/z_0))\right)^2}{k^2 u} \tag{6}$$

where u is the wind speed at reference height, z_{ref}, k is von Karman's 'constant', d is the displacement height and z_o is the roughness length. The parameters d and z_o are given explicitly as model parameters.

The actual evaporation from the canopy, E_a, is limited either by the amount of water in interception storage or by the potential evaporation rate. The interception storage is defined as the interception in the actual time step (S_{int}) plus the residual interception storage, and E_a given as

$$E_a = \min\left\{E_p, S_{\mathrm{int}} + \frac{S(t-1)}{\Delta t}\right\} \tag{7}$$

There is no reduction in E_a with the amount of water in storage and remaining intercepted water at the present time step is calculated as

$$S(t) = S(t-1) + (S_{\mathrm{int}} - E_a)\Delta t \tag{8}$$

Calculation of Interception Parameters

Following the method of Leyton *et al.* (1967), $S_{\max}$ is calculated from regression analysis of daily precipitation and throughfall measurement. When the actual amount of intercepted water on the canopy, S, is larger than $S_{\max}$ both direct throughfall and canopy drainage reach the ground. The shift of rainfall exceeding a lower limit, P_0, produces an inflection point in the P/T graph. This lower limit, P_0, is related to $S_{\max}$ by the following equation: $S_{\max} = P_0(1-p)$, where p is the free throughfall coefficient. Leyton *et al.* (1967) argue that the scatter of points to the right of the inflection point is caused mainly by variation in evaporation. Linear regression line should therefore be drawn through the data points with maximum throughfall, $T_{\max}$, as they represent data with minimal evaporation. Providing the stemflow is negligible (as assumed here, see Stand Description) this upper envelope would ideally be a line of unit slope: $T_{\max} = P - S_{\max}$ from which $S_{\max}$ is given from the intercept.

Site Description

The study site at Skogaby was located in the south-western part of Sweden, 30 km south-east of Halmstad (56°33.5' N, 13°13.5' E), about 16 km from the coast and at an altitude of approximately 100 m. a. s. l. The Skogaby project started in 1988 as a multidisciplinary ecosystem research activity. The experimental design was a randomised block design with four replicates. There were 30 plots, each with an area of 2,000 m^2. A detailed description of the experimental site is given by Bergholm *et al.* (1995) and Alavi (1996). Four plots were used to decrease the availability of water by artificial drought (D). A roof located 0.5-2.0 m above the ground prevented 2/3 of the throughfall from reaching the forest floor in half the plot, m^2, during the period from mid April to the end of September. These plots were denominated 8D, 11D, 16D and 23D.

Stand Description

The area was planted in 1966 with two provenance of Norway spruce (*Picea abies* (L.) Karst.), replacing the first generation of Scots pine (*Pinus sylvestris* L.) planted in 1913. The mean height of the studied plots (four drought and four control) in 1991 was 14.2 m and the mean stand basal area, sum of tree basal areas (m^2) divided by plot area (ha), for the same plots and year was 29 m^2 ha^{-1}. Corresponding values for Plot 23D in November 1990 (intensive measurement period) were 11.7 m and 35.5 m^2 ha^{-1}, respectively. The stands were completely closed and there was no understorey vegetation.

Nilsson and Wiklund (1993) measured leaf area index in the studied plots with a portable device (LI-COR LAI-2000). The device measured shoot area rather than

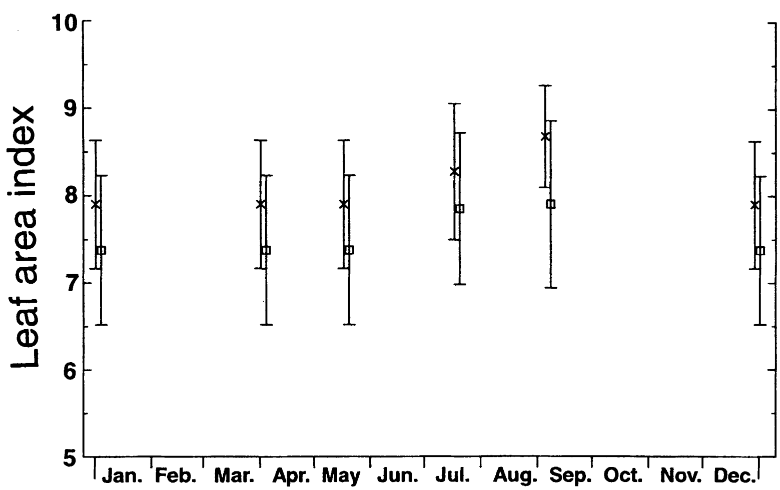

Fig. 1. The annual cycle of leaf area index in control (cross) and drought (squares) plots during 1991-1993: mean values and standard deviations, (data of Nilsson and Wiklund, 1993).

leaf area, so the measured values were multiplied by a correction factor of 1.60 according to Gower and Norman (1991) to obtain the actual *LAI*-values. Leaf area indices were generally high (Fig. 1). The spatial variation was about 2 units whereas the seasonal variation was less than 1 in the plots studied (Fig. 1).

Stemflow was assumed to be negligible because of the high *LAI*. Bergholm *et al.* (2001) measured stemflow in control plots during 16 months (Oct. 1990 to Feb. 1992) at Skogaby and found it to be about 1% of the throughfall.

Measurement of Throughfall

Throughfall was measured using water captured by roofs, intensively in 23.6×9 m^2 area of plot 23D during November 1990 and extensively in all D-plots during the periods from mid April to the end of September 1989-1991 and 1993.

The intensive measurement campaign was conducted using a trough gauge (roof ridge capping) with a collecting surface of 212.4 m^2. Data were recorded at 20 min intervals using a data logger.

The throughfall captured by roofs in D-plots was channelled into U-shaped PVC troughs and then fed to and recorded by a tipping bucket gauge during the extensive measurement period. The tipping bucket gauges were read once a week. However, the gauge in plot 23D was connected from June 20, 1989 to September 30, 1989 to a data logger which accumulated daily totals of throughfall. The volumes of water collected were converted to measurements of water depth, expressed in mm.

In addition to the measurements from the large roofs within the D-plots, we also measured the throughfall with funnel gauges in the control (C) plots; 3C, 15C, 18C and 24C. It was made using six randomly distributed funnel gauges in an 12.5×12.5 m area of each plot (45×45 m) which were measured and emptied at monthly intervals except for rainy periods when a weekly interval of measurement was applied. The funnel gauges had a diameter of 0.20 m (0.0314 m^2) and were installed at a height of 0.5 m above the soil surface. They were placed in isolated, dark boxes in order to prevent evaporation.

Measurements of Climate Parameters

Air temperature, air humidity, solar radiation and precipitation were measured at 1.5 m height and wind speed at 2.1 m height hourly in a 50×50 m gap about 300 m from the studied plots. During November 1990, the rain gauge was placed in an open area approximately 25 m north of plot 23D and precipitation was measured and accumulated on a 20 min time scale. The precipitation was measured by Swedish standard gauges (SMHI gauge). It was adjusted by +7% for the compensation of wind, wetting and evaporation losses (Seibert and Morén 1999). To estimate the wind speed at the top of canopy, the measured values were multiplied by a correction factor of 2.9 given by Alavi and Jansson (1995). This factor was obtained by comparing the measured values with the measurement of wind speed at 1 m above the canopy in the summer of 1991.

Estimation Procedure

1) S_{max} was estimated according to the method of Leyton *et al.* (1967) and using the daily data from summer 1989, Plot 23D (see section on measurement of throughfall).
2) The model was run using the estimated S_{max} and r_a about 1 s m^{-1} with 20-minute time step during the intensive measurement campaign (plot 23D, November 1990). The aerodynamic resistance was calculated assuming a value of 4 m for the roughness length and a value of 6 m for the distance from displacement height to the reference height. These values were estimated by Alavi and Jansson (1995) for the C- plots through calibration of SOIL with the help of the measured soil water potential.
3) The model was calibrated against the total measured throughfall during the intensive measurement period by adjusting i_{LAI} until the simulated throughfall was equal to the observed throughfall.
4) The model was tested separately for each of C and D plots using the estimated seasonal course of *LAI* which was assumed to be the same through all the studied years. It was run using daily time steps for the extensive measurement periods, from mid April to the end of September 1989-1991 and 1993.
5) A final adjustment of parameter values was further made when the agreement was not satisfactory (see following section on model acceptance) for the extensive measurement periods. In this case i_{LAI} was adjusted and a corresponding change of r_a was made to retain the agreement for the intensive measurement period.

Criteria for Model Acceptance

Two types of comparisons were made to identify the model performance:

1) Accumulated simulated throughfall was plotted against the corresponding measured values and simple linear regressions were computed with the measured values as independent variables. The slope of the regression equations was used as an indicator of model performance. Values of 1 would indicate exact agreement, less than 1 an underestimation, and larger than 1 an overestimation of throughfall.
2) Simulated total amount of throughfall in each plot and extensive measurement period were compared with measured values, and modelling efficiency (*EF*) and index of agreement (*d*) were obtained as follows

$$EF = \frac{\sum_{i=1}^{N}(O_i-\bar{O})^2 - \sum_{i=1}^{N}(P_i-O_i)^2}{\sum_{i=1}^{N}(O_i-\bar{O})^2} \tag{9}$$

$$d = 1 - \frac{\sum_{i=1}^{N} (P_i - O_i)^2}{\sum_{i=1}^{N} (|P_i - \bar{O}| + |O_i - \bar{O}|)^2} \quad (10)$$

where P_i is the predicted value; O_i is the observed value and $\bar{O}$ is the mean value of N observations. EF is an index of predictive performance and d indicates the degree to which predicted and observed values show similar deviation from the observed mean (Willmott 1981; Loague and Green 1991). If all predicted and observed values are the same then EF and d will be one (the maximum value). However, d varies between 0 and 1 whereas EF has no lower limit. In fact, any positive value of EF indicates an improvement over using the mean of the observations ($\bar{O}$) as the best estimator (Loague and Green 1991). The values of 0 for EF and 0.8 for d were chosen as the criteria for model acceptance.

Results

Climatic Conditions

The total precipitation was 104 mm for November 1990 and 481, 629, 655 and 671 mm for extensive measurement periods of 1989, 1990, 1991 and 1993 respectively. The growing season of 1989 was characterised by the smallest total precipitation and a long dry period during April-June with daily rainfall less than 10 mm. The years 1990 and 1991 were characterised by continuous wet weather. However, the rainfall distribution was more even during 1990 whereas daily rainfall of more than 20 mm occurred more often during 1991. The year 1991 was also distinguished by an extremely cold May and June. Finally, the year 1993 was characterised by a dry April and May but wet June, July and August including very large storms.

Throughfall – Precipitation Ratios

The ratio of throughfall to precipitation was generally low. D-plots (roof gauges) showed smaller ratio and less variability between plots and years than the data from C-plots (funnel gauges) (Fig. 2). The mean value was 0.50 for D-plots and 0.60 for C-plots. However, the ratio varied only between 0.41 to 0.53 for the roof gauges but between 0.42 to 0.71 for the data from funnel gauges.

Modelling and Estimation of Throughfall

1 – *Estimation of Interception Capacity According to the Method of Leyton et al. (1967)*

Daily totals of precipitation were plotted *versus* totals of throughfall in Plot 23D for

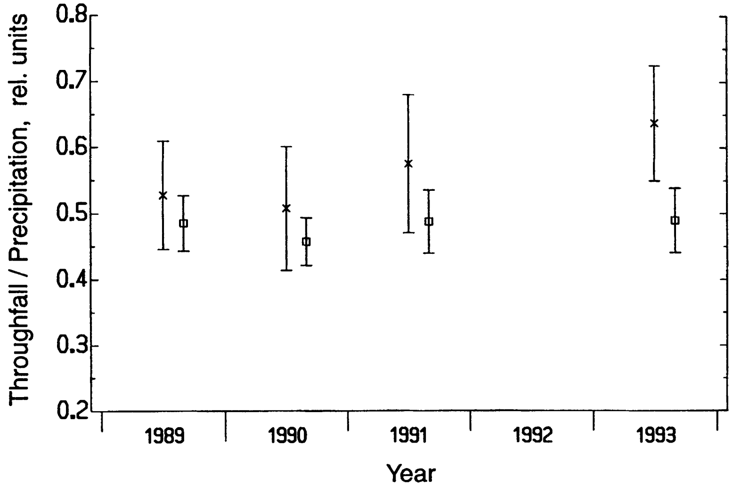

Fig. 2. Mean values and standard deviations of the ratio of throughfall to precipitation in plots with funnel (cross) and roof (squares)- gauges.

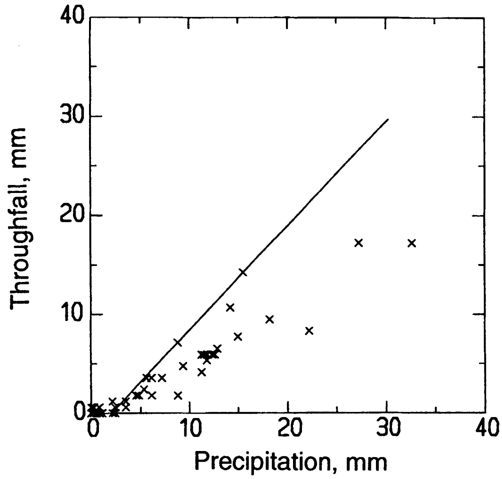

Fig. 3. Derivation of the saturation storage capacity S_{max} from precipitation and throughfall measured during summer 1989.

the period from June 20, 1989 to September 30, 1989 (Fig. 3). A S_{max} of 2 mm was calculated which corresponds to 0.23 mm on a leaf area basis (i_{LAI}). This value corresponds well with the values found by other researchers *e.g.*, Bringfelt and Håsmar (1974), Calder and Wright (1986) and Tallaksen *et al.* (1996). Comparable values for i_{LAI} have also been obtained in other studies. For example, values of 0.20, 0.23 and 0.24 mm were obtained by Aston (1979), Crockford and Richardson (1990) and Kelliher *et al.* (1992) in different stands of *Pinus radiata*, respectively.

Table 1 – Measured and modelled throughfall for different rainfall events during November 1990. Simulations were made with a 20-minute time step using three different parameter sets (I-III) and with a daily time step using parameter set III (III_{daily}).

		Throughfall (mm)				
			Simulated			
Date	Precipitation (mm)	Measured	I	II	III	III_{daily}
1-3/11	35.3	28.3	30	27	29.8	31.3
14-16/11	36	26	33.3	31.9	30.2	33
18-21/11	16.5	13.2	13.3	11.8	10	9.7
24/11	7.9	2.9	3	1.5	1.7	1.8
26-27/11	4.9	3.3	2.6	1.1	1.6	0.4
1-28/11	104	73.8	82.2	73.2	73.3	76.2

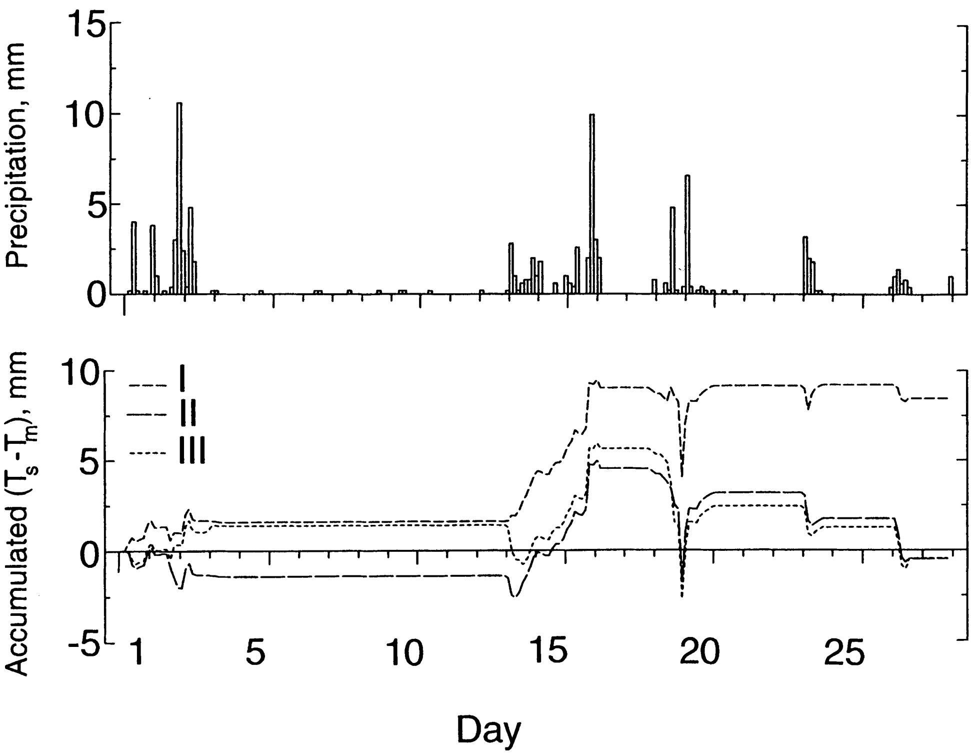

Fig. 4. Three-hour total precipitation and accumulated difference between simulated throughfall (T_s) and measured throughfall (T_m) for November 1990. Simulations were made using three different parameter sets (I-III).

2 – *First Parameterization of the Model, Parameter set (I)*

Two parameters are of great importance in simulation of interception losses: i_{LAI} which is the interception storage capacity per unit leaf area index and aerodynamic resistance, r_a. Using a 20-minute time step in the simulation and $i_{LAI} = 0.23$ mm, the total measured throughfall of 74 mm was overestimated by *at least* 11% (Table 1, Fig. 4). At least, because it was obtained through the least possible resistance (r_a about 1 s m^{-1}). The model overestimated the throughfall for all rain events except for the event with the smallest rainfall. It seems that the larger the storms, the greater the overestimation.

3 – *Second Parametrization of the Model, Parameter set (II)*

To overcome the problem of overestimation of throughfall, the value of i_{LAI} was increased to 0.4 mm. This is equivalent to a S_{max} of 3.5 mm. It resulted in a precise prediction of the total amount of throughfall (Table 1, Fig. 4). However, the model overestimated the throughfall by 23% during the largest storm (P=36 mm), and underestimated it by about 67% during the smallest storm (P =5 mm).

4 – *Test of Parameter set (II)*

Accumulated simulated throughfall was plotted against the corresponding measured values for the extensive measurement periods. The coefficient of determination, R^2, was larger for D-plots (roof gauges) whereas C-plots (funnel gauges) showed slope values closer to 1 (Figs. 5 and 6). The overall mean and standard deviation of the slopes was 1.23±0.2. This is an overestimation of the throughfall by 6 to 46%. *EF* and *d* were -1.03 and 0.62 indicating that the criteria for model acceptance, 0 and 0.8, were not fulfilled.

5a – *Third Parametrization, Parameter set (III)*

The overestimation of throughfall obtained indicated that i_{LAI} was too small. To avoid an underestimation of throughfall for the intensive measurement period, it was necessary to increase the resistance.

The aerodynamic resistance may be very uncertain as it was previously calculated based on rough estimates of z_o and *d* (Alavi and Jansson 1995). Consequently, the model value of z_o was decreased from 4 to 1.5 m and the i_{LAI} value was set to 0.7 mm (parameter set (III)) to obtain a good agreement (Table 1, Fig. 4). However, the throughfall was overestimated by about 5 to 15% during large storms (P=35 mm), and underestimated by 25-50% during medium and small storms ($p < 20$ mm).

5b – *Test of Parameter set (III)*

Tested on the extensive measurement periods, this parameter set resulted in larger R^2 and also slope values closer to 1 for D-plots compared to C-plots (Figs. 5 and 6). The overall mean and standard deviation of the slopes was 1.02±0.18. This is a prediction of the total seasonal throughfall to within 20% of the measured values. *EF* and

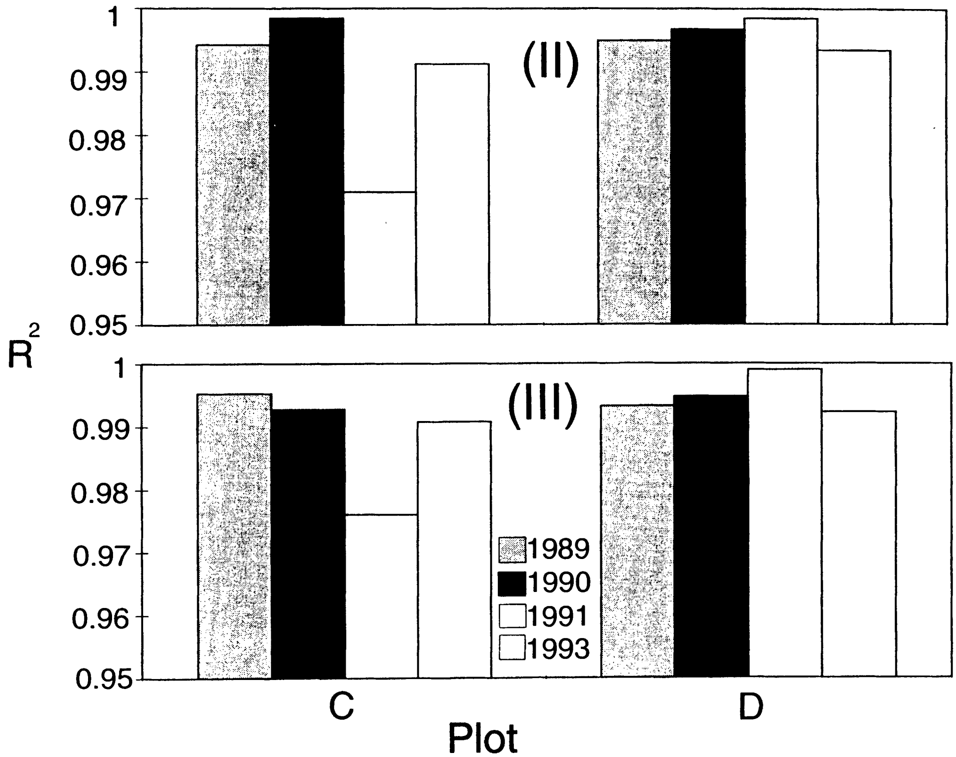

Fig. 5. The average values of coefficient of determination (R^2) for C- plots (funnel gauges) and D- plots (roof gauges) at each year from the simple linear regression of simulated throughfall on measured throughfall. Roman numbers represent different parameter sets.

d were now improved to 0.45 and 0.84 fulfilling the criteria for model acceptance. The 95% confidence intervals of the measured and simulated interception losses overlapped each other for all four extensive measurement periods (Fig. 7). However, the pattern of underestimation during small and medium intensities and overestimation during high intensities remained.

Discussion

D-plots showed lower throughfall as compared to C-plots despite a similar canopy structure. It may be explained by wetting and evaporation losses from the roofs in D-plots. D-plots showed also a less spatial and temporal variability which may be related to the larger collecting surface and the fact that the funnel gauges were not relocated during the measurement periods.

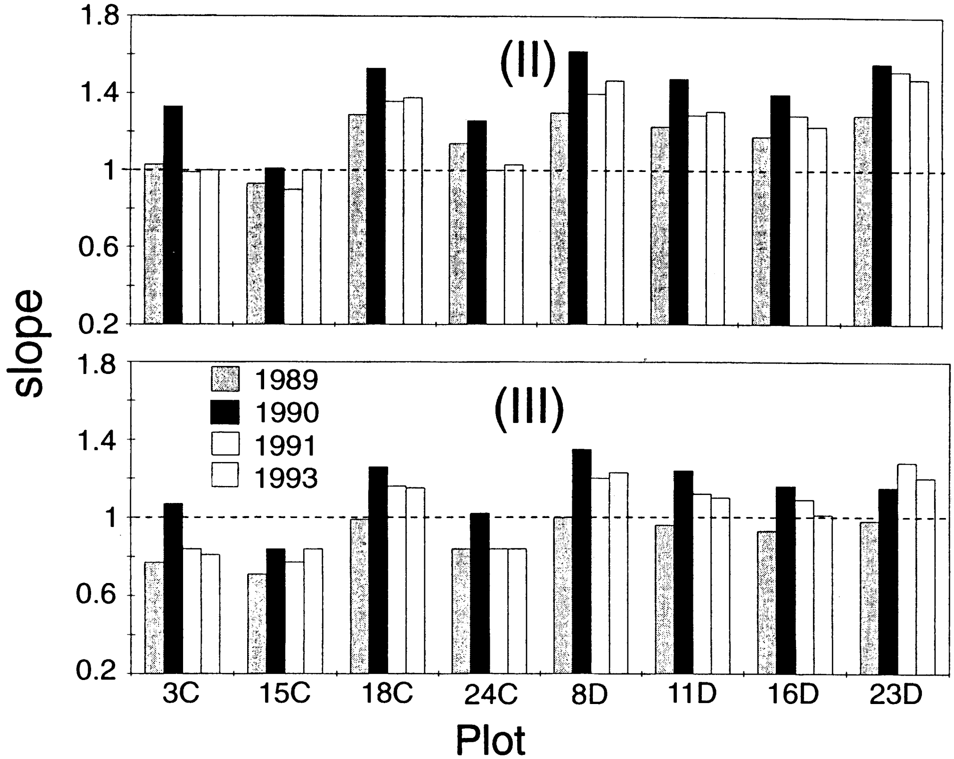

Fig. 6. The slope values from the simple linear regression of simulated throughfall on measured throughfall for different plots and years.

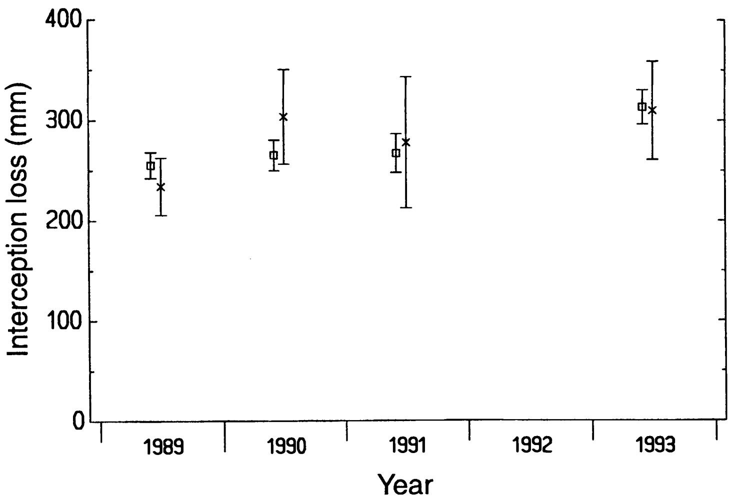

Fig. 7. Mean value with 95% confidence interval of observed (cross) and simulated (squares) total interception loss from both C and D- plots in each studied year.

The average measured total precipitation and interception losses for both C- and D plots during the extensive measurement periods were 610 and 280 mm, respectively. This shows that about 45% of the precipitation is lost as interception losses in the dense forest of Skogaby. However, the interception losses varied between 30 and 60% for different plots and years. Until now, there have been few reports showing such a large value for interception losses for coniferous forests (*cf.* Lundberg 1996).

Using the first parameter set, the model predicted well the throughfall for medium-sized storms, $5<P<20$ mm, whereas it slightly underestimated the throughfall for small storms, $P<5$ mm, and greatly overestimated it for large storms, $P>30$ mm (see Table 1, Fig. 4). In other words, the model went from overestimation to good estimation and underestimation of the interception evaporation as the rainstorm size increased. A similar tendency has been reported in other studies. Tallaksen *et al.* (1996) used measured throughfall in a coniferous forest stand to test the performance of three different types of models: the Nordic HBV model, the AMOR model and a simplified Rutter model. They found that the models underestimated the interception loss for storms exceeding 20 mm and overestimated it for storms with less than 10 mm rainfall. Jetten, (1996) used the Rutter model for prediction of interception loss from tropical rain forest and also found that the model underestimated the total interception loss. To increase the interception for large rainstorms, he extended the Rutter model with a layered representation of the canopy. This modification resulted in a considerable increase in the interception for most rainfall amounts compared with the Rutter model but still underestimated the loss from events with the largest precipitation. Also, Kelliher *et al.* (1992) obtained a lower modelled tree canopy interception during days when daily falls exceeded 20 mm. This weak point in many interception models could be related to the observed trend that interception loss rate during rain will increase with rainfall intensity (*e. g.* Stewart 1977; Pearce and Rowe 1980; Dunin *et al.* 1988). Dunin *et al.* (1988) speculated that the shattering of rain droplets high in kinetic energy produces a mist of small droplets causing high local concentrations of water plus vapour within the canopy space which may accelerate the turnover of water vapour during high intensity rainfalls due to an associated increase in the frequency and turbulent intensity of downdrafts. However, Klaassen *et al.* (1998) contrary to the statements above, recently reported that the water storage is the dominant process in interception of dense forest and that evaporation during rain is of minor importance. They tested different graphical methods, among them the Leyton method, using direct observations of water storage and evaporation and found that these methods underestimated water storage by a factor of 2 to 6 as compared to direct observations. They concluded that these methods are affected by a systematic error. The Leyton method probably underestimates the water storage because:

1. The data is chosen subjectively for regression line,
2. Rain drops may splash on an already wetted part of the canopy (Calder 1986),

3. Bark and undersides of leaves saturate slowly (Herwitz 1985),
4. Upper leaves shelter lower leaves.

Because the Leyton method has been used extensively with satisfying results, they recommended a decrease of the evaporation rate during rain to compensate the increase of S in order to yield realistic results. They suggest increasing of r_a as a useful tool to decrease evaporation.

We found fairly good agreement between modelled- and measured cumulative interception as we increased i_{LAI} and the resistance. We found also that increasing S_{max} was the main reason for the closer prediction of interception loss for large storms whereas increasing of resistance determined the agreement for the small storms. However, the model still underestimated the interception loss for large (>20 mm) storms and overestimated it for medium and small storms (<20 mm) like many other models. It is probably because different mechanisms govern water storage and drainage during different rain intensities. Water storage on bark might be an explanation. Bark is wetted slowly and an increase of storage with amount of precipitation could explain the overestimation of the interception loss for small storms and the underestimation for large storms. Herwitz (1985) showed that bark accounted for a large part of interception storage capacity. However, his interception studies were conducted in a tropical rainforest which has a much greater woody surface area than our temperate forest.

We approached the agreement for the extensive measurement periods mostly through the increase of the storage capacity. This indicates the major importance of the storage capacity in the long term modelling of interception loss. The model simulated between-year variation of throughfall better for D-plots. This is likely because of the large spatial variability associated with measured data from funnel gauges.

The estimated value of 0.7 mm for i_{LAI} corresponded to an interception storage capacity of 6 mm. This is three times larger than that obtained according to the method of Leyton *et al.* (1967). However, many possible errors may be involved in quantifying the interception process. One possible source of error is the use of different gauges for measurement of precipitation and throughfall. Using the same type of device for the measurements of throughfall and precipitation will decrease the error and make the results much more comparable. A number of reports have indicated that measurements of precipitation are the weakest link in estimating interception losses (Calder 1990; Neal *et al.* 1993). Other meteorological variables such as wind speed and relative humidity may also be affected by measurement errors. Another important source of error is the choice of daily time step in the modeling of throughfall for the extensive measurement periods. A test of the model showed that decreasing temporal resolution from 20 minute to a daily time step does not change the general pattern discussed above but amplifies the differences (Table 1). Using a daily time step, contributes to the underestimation of throughfall for small events because of the "Smearing effect". Smearing the small storms over 24 hours results in a smaller intensity than daily interception capacity, leading to less simulated throughfall than actually observed during a short event. In contrary, more throughfall is simulated for large events as the potential evaporation is smeared over the whole day leading to a lower evaporation rate.

Conclusions

This study showed that between 30 to 60% of the precipitation is lost as interception loss from the dense and young spruce stands at Skogaby. Tree canopy water storage capacity on a leaf area basis was estimated to 0.7 mm which was three times larger than that obtained graphically. The use of the graphical method of Leyton *et al.* (1967) resulted in underestimation of the canopy storage capacity. However, many possible errors, particularly in measurement of precipitation and throughfall, may be involved in quantifying interception losses.

This study demonstrated that a simple model is able to estimate fairly well the long-term cumulative interception loss from a dense spruce forest using two important parameters, the interception storage capacity per unit *LAI* and an estimate of aerodynamic resistance. It was concluded that the storage capacity was of major importance in modelling of the long-term interception loss from dense forest.

The simulated throughfall values varied less than the measured values, indicating a lack of sensitivity of the model to adequately describe the interception process for individual storm events.

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