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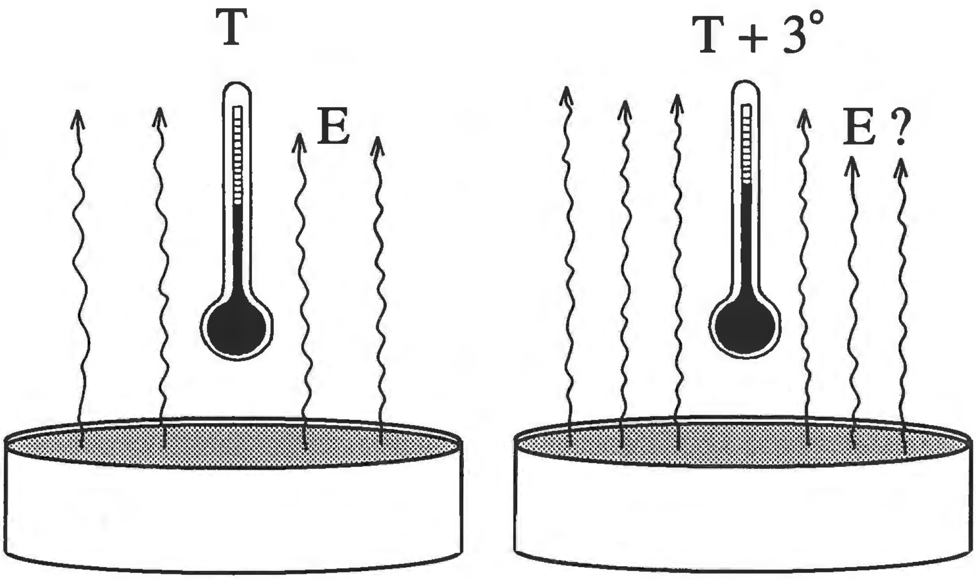

CONCEPTUAL MODELLING OF EVAPOTRANSPIRATION FOR SIMULATIONS OF CLIMATE CHANGE EFFECTS

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Abstract

The evapotranspiration routines in existing conceptual hydrological models have been identified as one of the weaknesses which appear when these models are used for the simulation of hydrological effects of a changing climate. The hydrological models in operational use today usually have a very superficial description of evapotranspiration. They have, nevertheless, been able to yield reasonable results. The objective of this paper is to analyse and suggest modifications of existing evapotranspiration routines in conceptual hydrological models to make them more appropriate for use in simulation of the effects of a changing climate on water resources.

The following modifications of the evapotranspiration routine were formulated and tested in the HBV model: Temperature anomaly correction of evapotranspiration, potential evapotranspiration by a simplified Thornthwaite type formula, interception submodel, spatially distributed evapotranspiration routine and alternative formulations of lake evapotranspiration. Sensitivity analyses were thereafter made to illustrate the effects of uncertainty in the hydrological model structure versus those of the uncertainty in the climate change predictions.

Key words
Hydrological modelling, evapotranspiration, interception, climate change

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1. INTRODUCTION

Evapotranspiration is a key process in the simulation of hydrological effects of a changing climate. This question was the subject of a seminar held at Vettre in Norway in March 1992 (Tallaksen and Hassel, 1992). The modelling of evapotranspiration losses is not trivial in this respect. There are a number of factors which should be considered, for example vegetation response to a changing climate. This effect is probably small compared to the effect of active forest management. It therefore seems reasonable to concentrate on the climatic scenario and to treat vegetation feed-backs as a secondary problem.

The hydrological models in operational use today often have a very simplified description of evapotranspiration. They have, nevertheless, been able to give satisfactory simulations of discharge. As an example, the HBV model (Bergström, 1976 or 1992), which is the basis for hydrological forecasting at the Swedish Meteorological and Hydrological Institute, does not account explicitly for precipitation measurement errors nor interception losses. The crude evapotranspiration routine in the model has sometimes led to the use of unrealistically low snowfall correction factors, even well below unity (see e.g. Bergström, 1990). It is therefore not suited for detailed analyses of vegetation feed-backs, although there are some empirical studies where it has been used in this respect (e.g. Brandt, 1990).

For the use in climate change simulations, it is an advantage if the evapotranspiration module can be kept simple, without dependence on additional input data, such as e.g. radiation and wind conditions. Often, only predicted changes in precipitation and temperature are at hand. One alternative to the Penman standard values is the Priestly-Taylor equation (Priestly and Taylor, 1972). Evremar (1994) tested this equation in the HBV model for some basins in northern Sweden, without any significant improvements of discharge simulations as compared to the original model. A disadvantage of using the Priestly-Taylor equation in climate change simulations is the need for radiation predictions, in addition to precipitation and temperature.

2. OBJECTIVE

The objective of this paper is to analyse and suggest possible modifications of existing evapotranspiration routines in conceptual hydrological models to make them more appropriate for use as simulation tools for the analysis of the effect of a changing climate on water resources. Sensitivity analyses are thereafter made to illustrate the uncertainty due to model structure versus the uncertainty due to the climate change predictions.

3. MODEL DESCRIPTION AND DATA BASE

The starting point of the modelling work was the HBV model as described by Bergström (1976 and 1992). The model is a conceptual description of the main elements of the hydrological cycle, such as precipitation, snow accumulation and melt, infiltration, evapotranspiration, percolation and runoff. Long term mean values, normally with a monthly resolution, of potential evapotranspiration, E_p are used as input to the model. The estimates by Wallén (1966) and Eriksson (1981) of E_p according to the Penman formula are normally used in standard applications in Sweden.

All simulations were made within the framework of the newly developed Integrated Hydrological Modelling System, IHMS. A daily time step was used in the precipitation and temperature input. All tested models were calibrated using the process oriented calibration scheme developed by Harlin (1991), and all comparisons were made between computed and recorded discharge. The automatic calibration was also complemented by visual inspections of the model simulations. The model performance, i.e., the agreement between computed and measured discharge, was measured by the commonly used efficiency criterion R^2, as suggested by Nash and Sutcliffe (1970).

In Sweden, the model is usually run with constant monthly values of Penman evaporation as input. For this study, a linear interpolation was used between the monthly estimates. This simple modification generally improved the simulation results slightly.

The following tests were made:

3.1 Temperature anomaly correction of evapotranspiration (ETF)

Lindström and Bergström (1992) reached some promising simulation results for the relatively cold summers of 1987 and 1991, by taking into account the actual temperature deviation (Figure 1) from the long term monthly mean temperature. The temperature anomaly is here intended as an index of weather type, primarily of radiation conditions. In the present study mean temperatures were calculated for each day of the year.

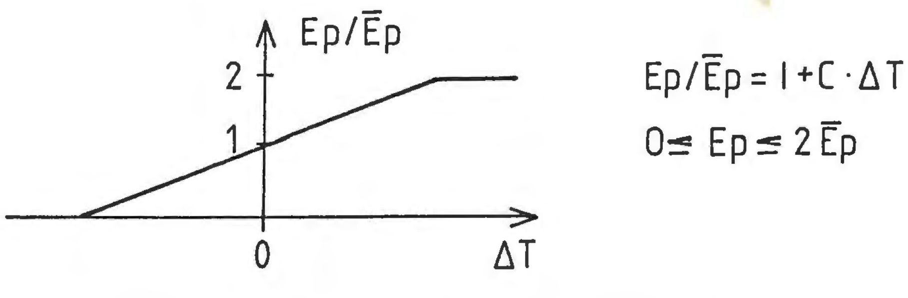

Figure 1. *The temperature anomaly correction of potential evapotranspiration. $\bar{E}_p$ = long term mean potential evapotranspiration, E_p = current potential evapotranspiration, ΔT = temperature deviation from long term mean, C = a model parameter, called ETF in the text.*

3.2 Potential evapotranspiration by the Thornthwaite formula (THO)

A simplification of the Thornthwaite formula (see e.g. Rosenberg et al., 1983) was tested, in which the evapotranspiration was calculated only as a linear function of temperature above the freezing-point:

$$E_p = K_T \cdot STF(t) \cdot T$$

The parameter K_T was calibrated and STF is a seasonally varying coefficient (Table 1) as used by Saelthun et al. (1990). The potential evapotranspiration here becomes a function of altitude, since the model computes the temperature in each elevation zone according to a temperature lapse rate, normally of about 0.6 °C per 100 m.

Table 1. Monthly factor STF in the Thornthwaite type computation of potential evapotranspiration.

J	F	M	A	M	J	J	A	S	O	N	D
0.7	0.7	0.8	1.0	1.3	1.2	1.1	1.0	0.9	0.8	0.7	0.7

A reduction of the evapotranspiration during rainy days was tested in combination with this test. The evapotranspiration was reduced by a factor exp(-kP), where k is a model parameter and P is the precipitation. This reduction did not lead to any improvements and was therefore not used in any of the following simulations.

3.3 Precipitation corrections and interception submodel (INT)

The fact that interception losses were not explicitly taken into account earlier was partly compensated by the use of uncorrected precipitation measurements. Correction factors for wind losses, wetting and evaporation for Sweden were suggested by Eriksson (1983). He suggested different wind factors for rain and snowfall, about 5 - 15 % for rain and much higher losses for snowfall. The high losses suggested by Eriksson have later been questioned. For example, Carlsson (1985) found much smaller wind losses for snowfall, approximately 10 - 20 % losses on average over a year (Figure 2). A compromise was made for the present study. The aerodynamic losses for rain were taken from Eriksson (1983) whereas the loss for snowfall was calibrated, but with unity as the lower accepted boundary. An advantage of using Eriksson's values as the starting point is that he made a classification of the exposure of all the precipitation gauges operated by the SMHI. Individual correction factors can thereby be used for individual stations.

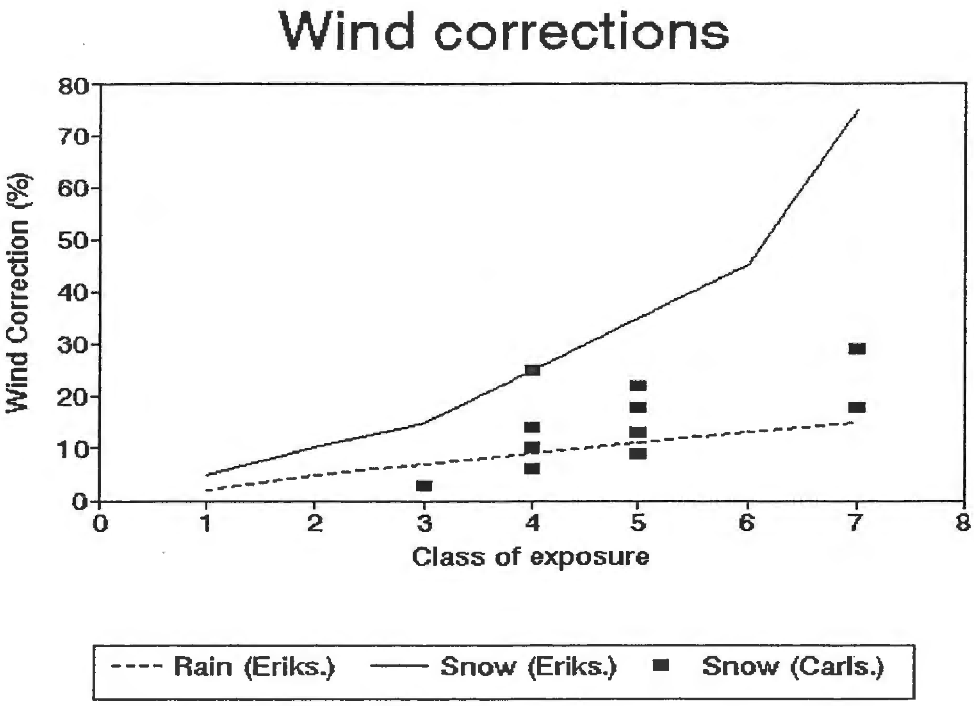

Figure 2. *Wind corrections for the SMHI precipitation gauge versus class of exposure (according to Eriksson, 1983), based on the results by Eriksson (1983) and Carlsson (1985).*

The importance of interception losses has been pointed out by for example Gash and Stewart (1977), Bringfelt and Lindroth (1987) and Lindroth (1993). Standard Penman values as normally used in the HBV model are not representative for forests, a fact which was actually also stressed by Eriksson (1981). Detailed studies and measurements of the interception losses from Swedish forests were carried out by Lundberg (1993). She concluded that interception losses can be considerable even during the winter.

A very simple storage model was tested in this study, based on a maximum storage capacity, ICMAX (Figure 3). No difference was made between snow and rain, and ICMAX was assumed constant during the year. The storage capacity was set to 2 mm in forests (cf. Bringfelt, 1982 or Lundberg, 1993) and neglected in open areas. The evaporation, E_I, from the interception storage was assumed to be potential, E_P. The transpiration from the vegetation can be assumed to cease while the canopy is wet (Bringfelt, 1982). In the original HBV model, the transpiration from the soil, E_{SM}, is computed as a function of the potential evapotranspiration and the soil moisture storage SM (see e.g. Bergström, 1992). In combination with the interception routine, an upper limit was here assumed for the transpiration, E_T, to account for the time during which there is evaporation of intercepted water. The transpiration, E_T, was set equal to E_{SM}, as long as this did not exceed an upper limit of:

$$E_T = \frac{E_{max} + E_{SM}}{2}$$

where $E_{max} = E_p - E_I$.

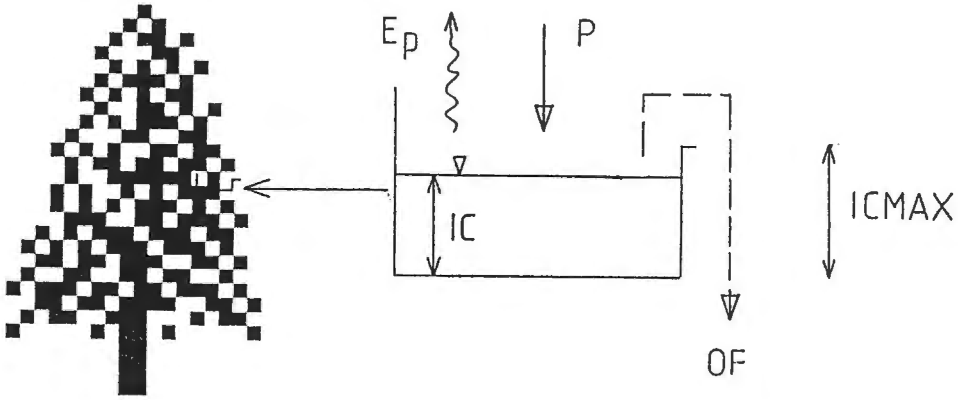

Figure 3. The interception routine. P = precipitation, OF = immediate overflow which occurs as soon as the storage, IC, reaches the storage capacity, ICMAX. No drainage takes place while IC < ICMAX.

The behaviour of the interception routine was reasonable. Out of the total modelled evapotranspiration losses from land areas in the Torsebro basin, about 30 % stemmed from interception losses, and the rest from transpiration. These values are in the same order of magnitude as those reported by Bringfelt for Velen, where interception losses were estimated to be about 20 % of the total evapotranspiration. The runoff from forested areas in the Torsebro basin was about 80 mm/year lower than for open areas, a figure which corresponds to estimates of the difference between runoff from forests and open areas (Johansson, 1993) in the southernmost province of Sweden, Skåne, where part of the basin is located.

3.4 Spatially distributed evapotranspiration routine (DIS)

The evapotranspiration process in the original HBV model is more or less lumped, with only a separation between lakes and land. As an example, the potential evapotranspiration does not depend on altitude. Neither does the model take into account the spatial distribution of snow within elevation zones. Instead, evapotranspiration is assumed to be negligible as soon as the ground is covered by snow. The low evaporation from snow is supported by studies by Bengtsson (1980), Lemmelä and Kuusisto (1974) and Vehviläinen (1992). Snow evaporation was therefore neglected in this study.

However, snow does neither accumulate nor melt evenly over a basin. The snow distribution within elevation zones was therefore described by a division into snow classes, with a difference in accumulation rate. Three classes of equal size were used, and the snow accumulation was assumed to vary linearly between the snow classes (Figure 4). The coefficient of variance in the snow accumulation rate was estimated, depending on vegetation and geographic conditions, by comparison with the values on snow cover variability which were reported by Gottschalk and Jutman (1979).

By water balance calculations, Evremar (1994) found a decrease in evapotranspiration of some 20 - 50 mm per 100 m and year in the Swedish mountain range. This corresponds to roughly 10 % per 100 m, a value which was here used as a lapse rate for potential evapotranspiration.

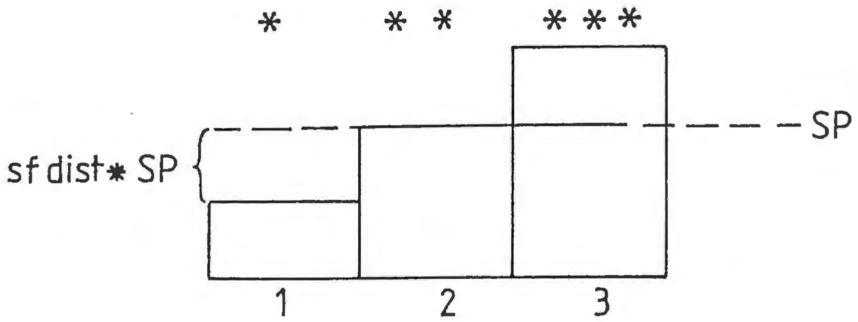

Figure 4. An example of the snowfall distribution routine with 3 snow bands.

3.5 Alternative formulations of lake evaporation (LAK)

The alternative evapotranspiration routine tested by Lindström and Bergström (1992) gave the largest improvements in some areas with a high lake percentage. Lake evaporation was studied further here, since lakes represent a major sink term during dry summers. Wallén (1966) suggested that his estimates of potential evapotranspiration for soils should be increased by 40 % for open water, mainly due to differences in albedo. This factor has been used in most applications in Sweden, even after the introduction of the values by Eriksson (1981). In practice, however, it has been noted that this assumption sometimes leads to problems with simulations in basins with a high lake percentage. It is felt that lake evaporation is overestimated. Eriksson used a lower albedo for the soil surface and his evaporation values are about 15 % higher than Wallén's. A more appropriate correction factor for Eriksson's values should be around 1.1. This value was also found by Kristensen (1979). The factor 1.4 was consequently used for Wallén's values and 1.1 was used for Eriksson's values in the modified lake evaporation routine.

A simple time lag of one month was introduced in the evapotranspiration from the lake surface for this study. The maximum lake evaporation therefore occurs one month later than the corresponding maximum from land areas. This is reasonable, since e.g. Eriksson (1981) disregarded the heat storage in the soil or in lakes in his estimates of Penman E_p. Observations of lake evaporation in Finland generally point to July as being the month with the highest values (e.g. Järvinen, 1978), whereas the potential evaporation, at least in Sweden, according to the Penman equation usually culminates in June. An independent check was made by using data compiled by Jutman (1975) on observations of lake evaporation in Sweden (Figure 5).

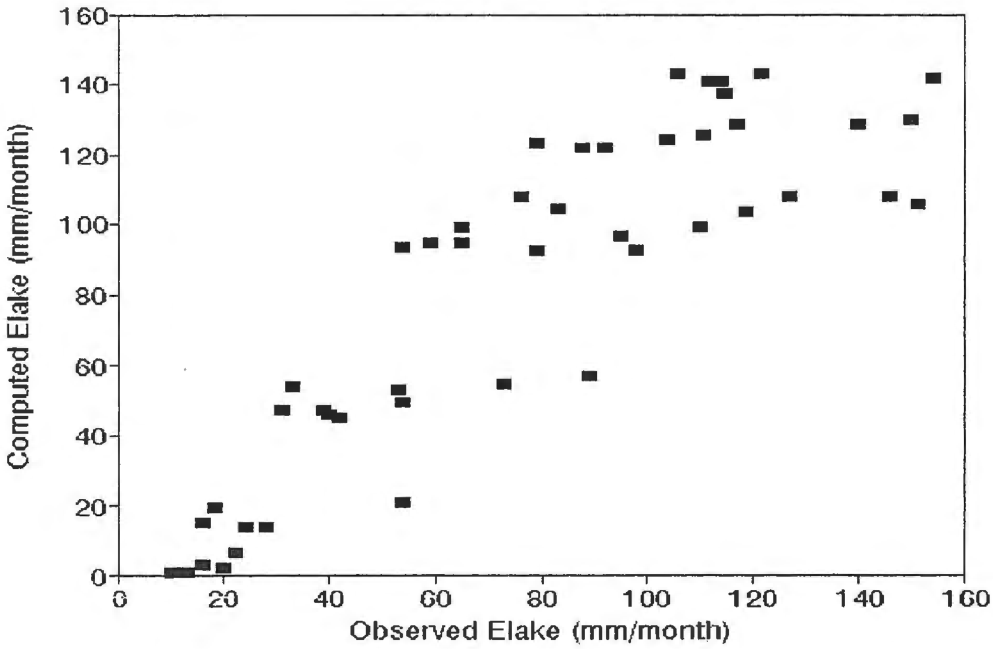

Figure 5. *Independent check of the modelled lake evaporation compared with observations compiled by Jutman (1975). Monthly data from individual lakes, mostly in southern Sweden, compared with estimates by Eriksson (1981) multiplied by 1.1 and delayed one month.*

A simplification in the ordinary HBV model is that the lakes are assumed to be covered by ice from 1 January to a fixed date in spring, a period during which there is no evaporation. These fixed dates do not respond to a climate change, and a different approach was therefore adopted. The lake surface temperature, T_L, was calculated as

$$T_L(t) = (1-k) \cdot T_L(t-1) + k \cdot T_A(t)$$

where T_A is the air temperature. The parameter k was set so that the corresponding time constant, i.e. 1/k, became equal to 30 days. In principle, this parameter could be allowed to depend on the size and geometry of the lake. When the surface temperature went below freezing, an ice cover was assumed, and evaporation was set equal to zero. In the spring, evaporation was resumed when T_L exceeded freezing. A similar method for predicting sea ice formation from air temperature was presented by Rodhe (1955). The simple ice model was here tested with independent data on the ice break-up dates for the lakes Kultsjön and Ivösjön, the latter near Torsebro. The agreement was good, considering the simplicity of the model (Figure 6).

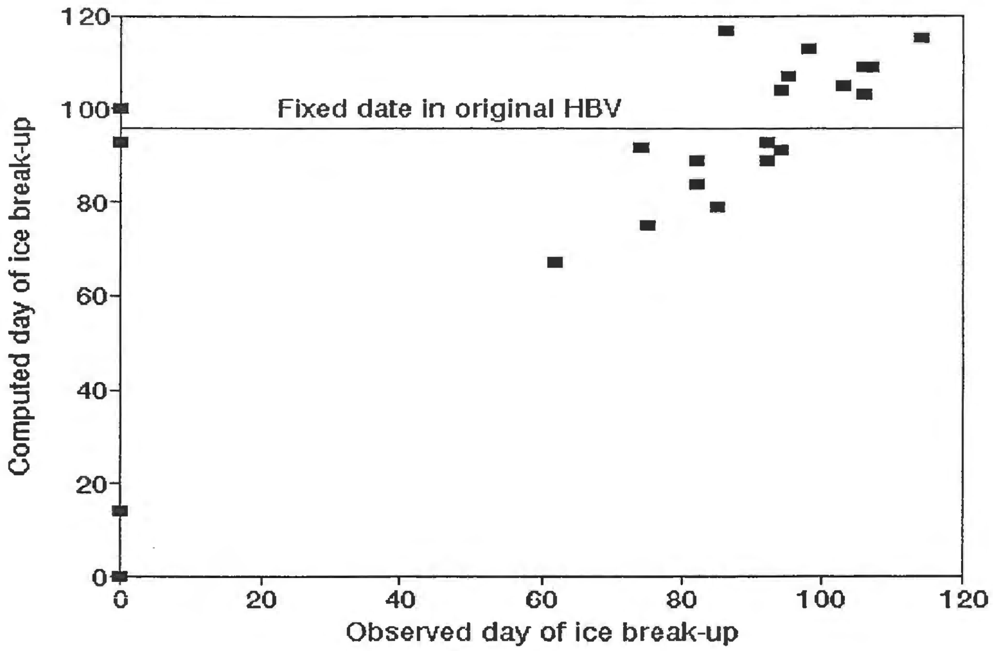

Figure 6. *Observed ice break-up days for lake Ivösjön compared to those computed by the model for Torsebro. Day 1 is 1 January and so on.*

3.6 Full model with ETF (F-ETF)

The temperature anomaly model, ETF, was combined with the models of the preceding tests 3.3, 3.4 and 3.5 (INT, DIS and LAK).

3.7 Full model with Thornthwaite equation (F-THO)

The Thornthwaite type model, THO, was combined with the models of the preceding tests 3.3, 3.4 and 3.5 (INT, DIS and LAK). However, no lag between lake evaporation and the potential evaporation was used in this test, since there is already a lag between potential evaporation and air temperature. Neither was the lapse rate of potential evaporation with altitude used, since the temperature decrease at high elevations automatically results in a lower evaporation.

3.8 Data base

Basins with a good data coverage were selected to cover the most important physiographic conditions in the Nordic countries (Figure 7). Key characteristics and simulation periods for the basins are summarized in Table 2.

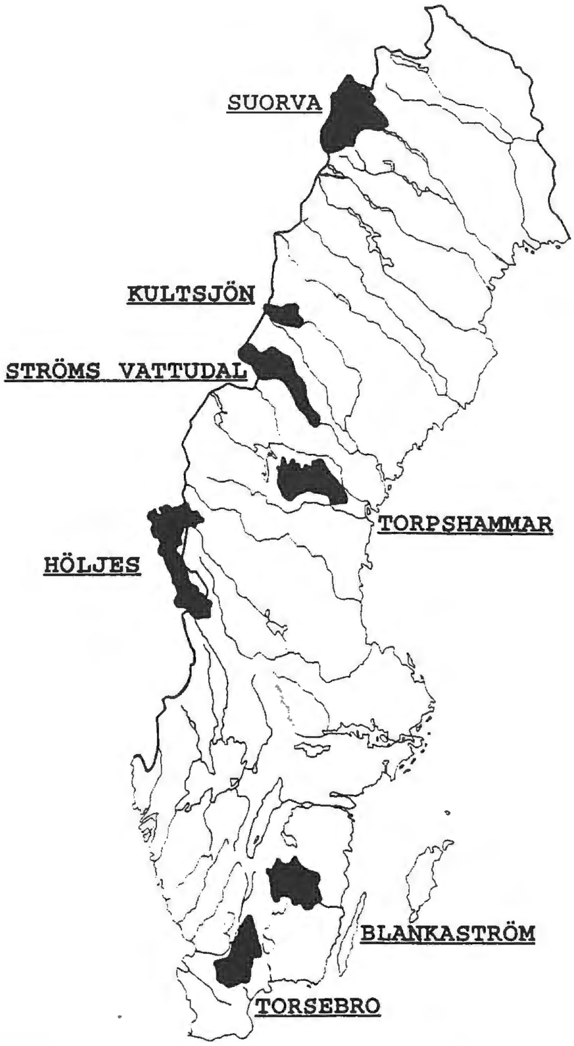

Figure 7. *Geographical location of the basins under study.*

Table 2. *Data base and calibration and verification periods. All computations were made from 1 September to 31 August.*

Basin	Area (km^2)	Lake area (%)	Forest area (%)	Calibration period	Verification period
Suorva	4688	14	3	1979-89	1969-79
Kultsjön	1109	7	28	1969-79	1979-89
Ströms Vattudal	3851	12	67	1979-89	1969-79
Torpshammar	4291	10	74	1979-89	1969-79
Blankaström	3446	7	72	1979-89	1969-79
Torsebro	3676	9	59	1969-79	1979-89
Höljes	5975	8	46	1969-79	1979-89

4. CLIMATE SENSITIVITY ANALYSIS

Sensitivity analyses were made to illustrate the uncertainty due to model structure versus the uncertainty in the climate change predictions, with the reference model and the two full models F-ETF and F-THO. Apart from the present climate, two scenarios were simulated, with temperature changes of +3 and -3 °C for each month over the whole simulation period. These values agree well with the recommendations by SILMU (1993). The period 1 September, 1969, to 31 August, 1989, was used in all simulations. The temperature was increased without affecting the precipitation amounts but only the fraction of rainfall and snowfall. The parameter ETF varied considerably between basins. A regional value of 0.15 was therefore used in the climate sensitivity analysis. Additional sensitivity runs were made for Höljes. The parameters K_T and ETF for all basins were grouped together, and the standard deviations were computed. The obtained values of the 2 parameters were thereafter increased by one standard deviation and the models were recalibrated, before running the climate sensitivity analysis again.

5. RESULTS

The differences in model agreement were rather small between the different test model versions (Table 3). Differences in the simulated runoff were nevertheless clearly visible. Examples of output from some different model versions are shown in Figures 8 and 9.

Table 3. R^2 in % for the reference model, and the difference in R^2 from the reference simulation for the other tests (calibration and verification periods).

Basin	Ref	ETF	THO	INT	DIS	LAK	F-ETF	F-THO
Suorva cal	90	0	0	0	0	0	+1	0
Kultsjön cal	87	0	0	0	+1	0	+1	+1
Ströms Vattudal cal	87	0	-1	-3	-1	0	-3	-3
Torpshammar cal	90	+1	+1	-1	0	0	-2	-1
Blankaström cal	83	+1	0	0	+1	-1	-1	-6
Torsebro cal	92	+1	0	+1	+1	0	+1	-2
Höljes cal	91	+1	0	0	+1	-1	+1	-1
Suorva ver	76	0	0	+1	+1	0	+2	+2
Kultsjön ver	81	0	0	-1	+1	0	0	0
Ströms Vattudal ver	78	0	0	-5	0	0	-4	-4
Torpshammar ver	87	-1	+1	0	+2	+1	-2	-1
Blankaström ver	87	0	-3	0	0	-2	-1	-1
Torsebro ver	86	+2	+1	-1	+1	0	+1	-2
Höljes ver	89	0	-1	-1	0	-1	-1	-3

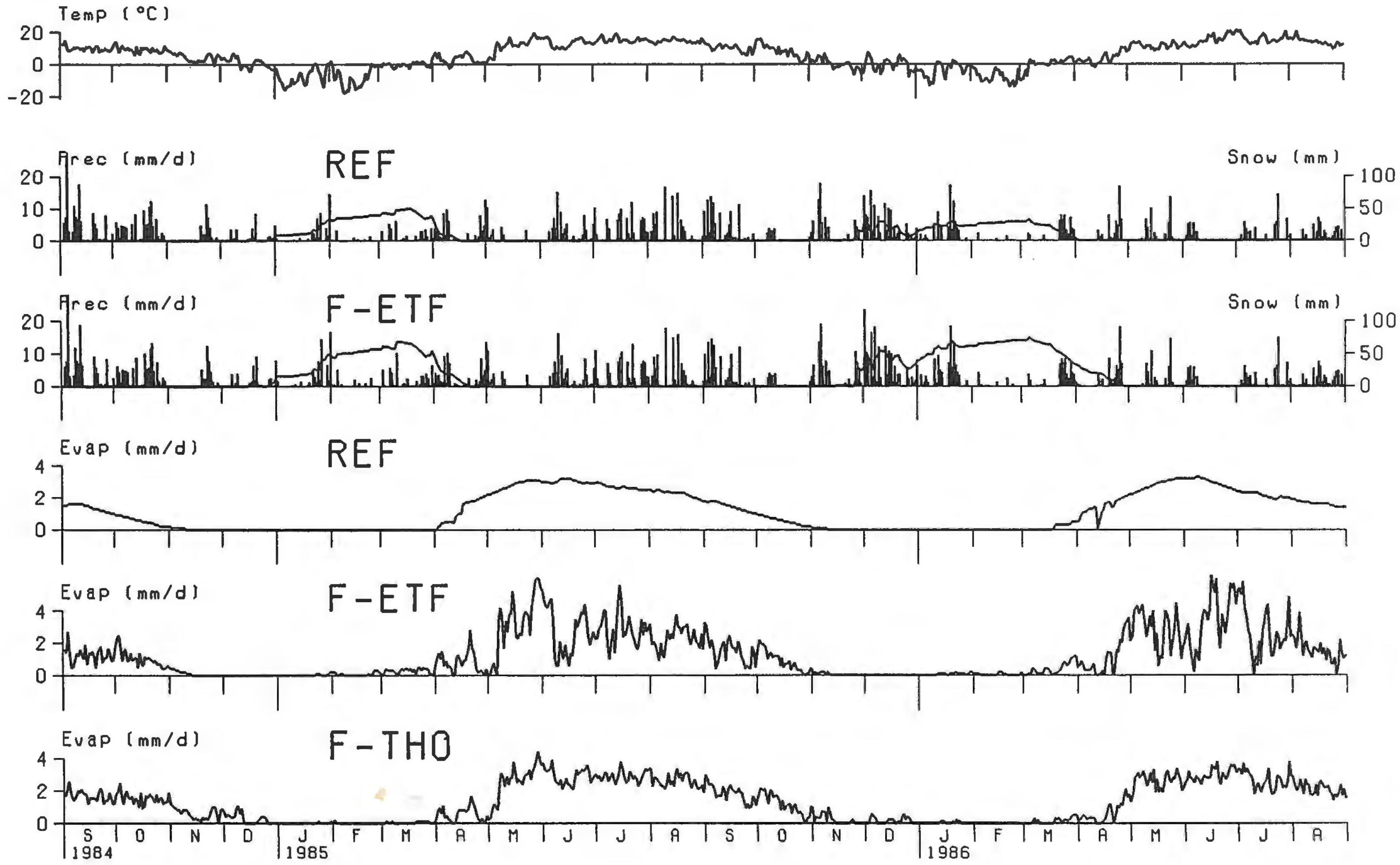

Figure 8. *Example of model output from the verification period for Torsebro. From top to bottom: temperature, precipitation and snow in the REF and F-ETF models respectively, evapotranspiration in the REF, F-ETF and F-THO models respectively. Note that the precipitation and snowpack of the F-THO model are very similar to those of the F-ETF model.*

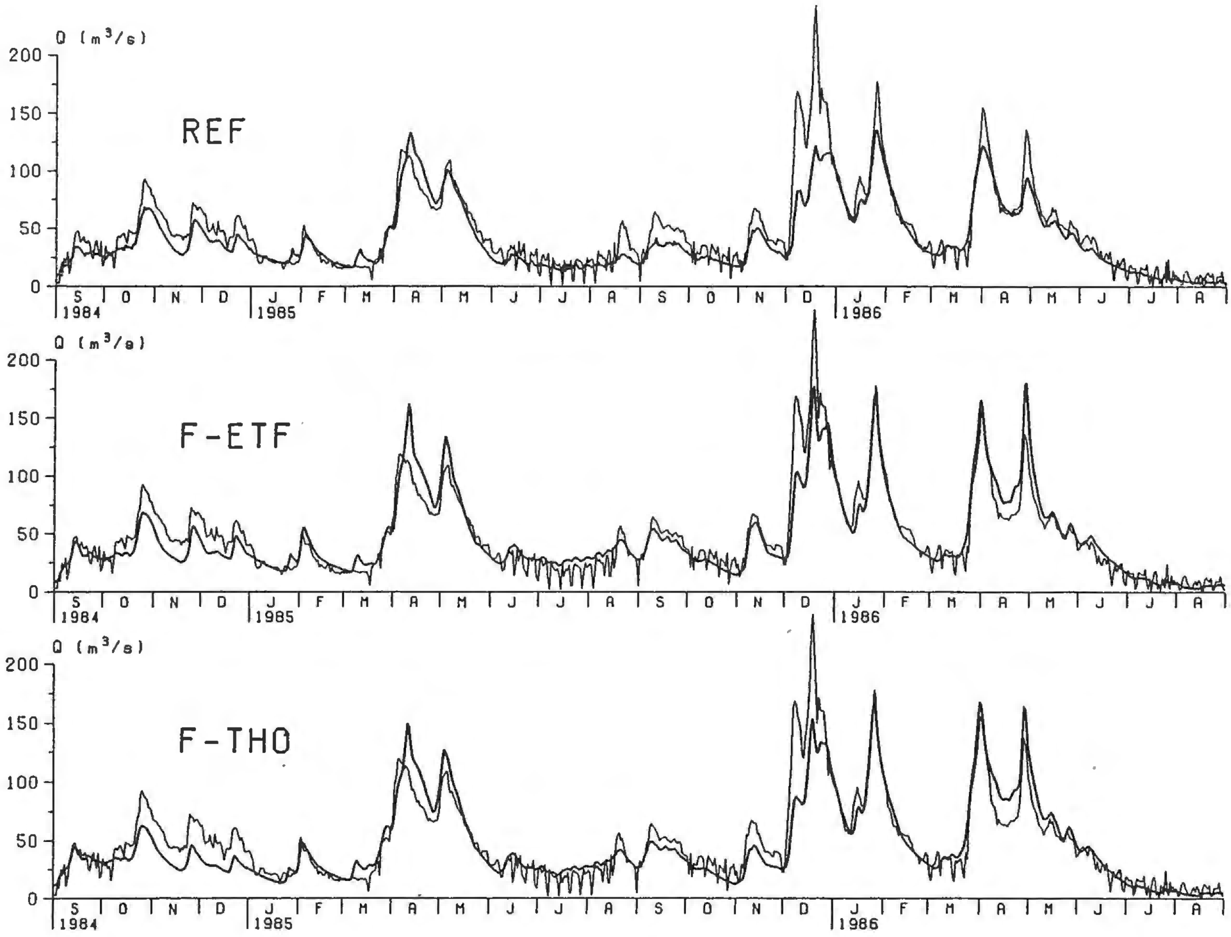

Figure 9. *Example of model performance from the verification period for Torsebro. From top to bottom: Simulated (thick line) and observed discharge (thin line) by the REF, F-ETF and F-THO models respectively.*

In general, the introductions of ETF and snow distribution improved the results slightly. On the other hand, the interception model usually gave poorer results than the original model, particularly when used without any of the temperature-dependent evapotranspiration models. The snowfall correction factors (Table 4) were nevertheless more realistic (cf. Figure 2), since the evapotranspiration increased. Brandt (1986) tested a slightly different interception routine in the PULSE model, which is similar to the HBV model. As in the present study, she arrived at more realistic snowfall correction factors, but with slight deteriorations in model performance. The parameter K_T was found to vary less between basins than ETF (Table 4).

Table 4. Some parameter values obtained by the calibration and correction factors for rainfall, crain, and snowfall, csnow, used in the simulations.

Basin	Original model		F-THO		K_T (THO, mm °C^{-1} day^{-1})	ETF (C)
	crain	csnow	crain	csnow		
Suorva	1.00	0.91	1.10	1.19	0.25	0.10
Kultsjön	1.00	1.26	1.10	1.36	0.21	0.15
Ströms Vattudal	1.00	1.16	1.09	1.39	0.21	0.10
Torpshammar	1.00	0.75	1.07	1.07	0.22	0.15
Blankaström	1.00	0.91	1.08	1.18	0.19	0.10
Torsebro	1.00	0.72	1.07	1.10	0.17	0.20
Höljes	1.00	1.11	1.09	1.31	0.22	0.10

The results from the sensitivity analysis are given in Table 5 and Figure 10. Little difference was found in the total runoff volumes between the prediction results using the two different models in the northern basins. In the south, however, the results were found to depend more on the choice of model, because of the greater importance of evaporation in the water balance. The increase of model parameters ETF and K_T with one standard deviation had little effect on the simulated volume. The changes were moderate and the models were recalibrated after the increase, which means that other parameters compensate for the increase in evaporation. Figure 11 shows the result for Höljes after an increase of K_T.

Table 5. Results from the sensitivity analysis of climate change predictions. The deviations are given in percent between the runoff simulated according to the present climate, and the runoff simulated after the climate change. No meaningful results were obtained when the temperature was decreased in Suorva, due to build-up of glaciers, which were not included in the model.

Basin	ΔTemp = -3 °C			ΔTemp = +3 °C		
	Ref	F-ETF	F-THO	Ref	F-ETF	F-THO
Suorva	-	-	-	-1	-1	-4
Kultsjön	+5	+11	+10	-5	-14	-14
Ströms Vattudal	+8	+26	+20	-5	-19	-20
Torpshammar	+8	+45	+29	-6	-30	-26
Blankaström	+15	+79	+46	-5	-41	-40
Torsebro	+15	+66	+40	-5	-38	-35
Höljes	+7	+28	+25	-5	-25	-23
Höljes (+1 st.dev. in K_T and ETF)	-	-	-	-	-23	-23

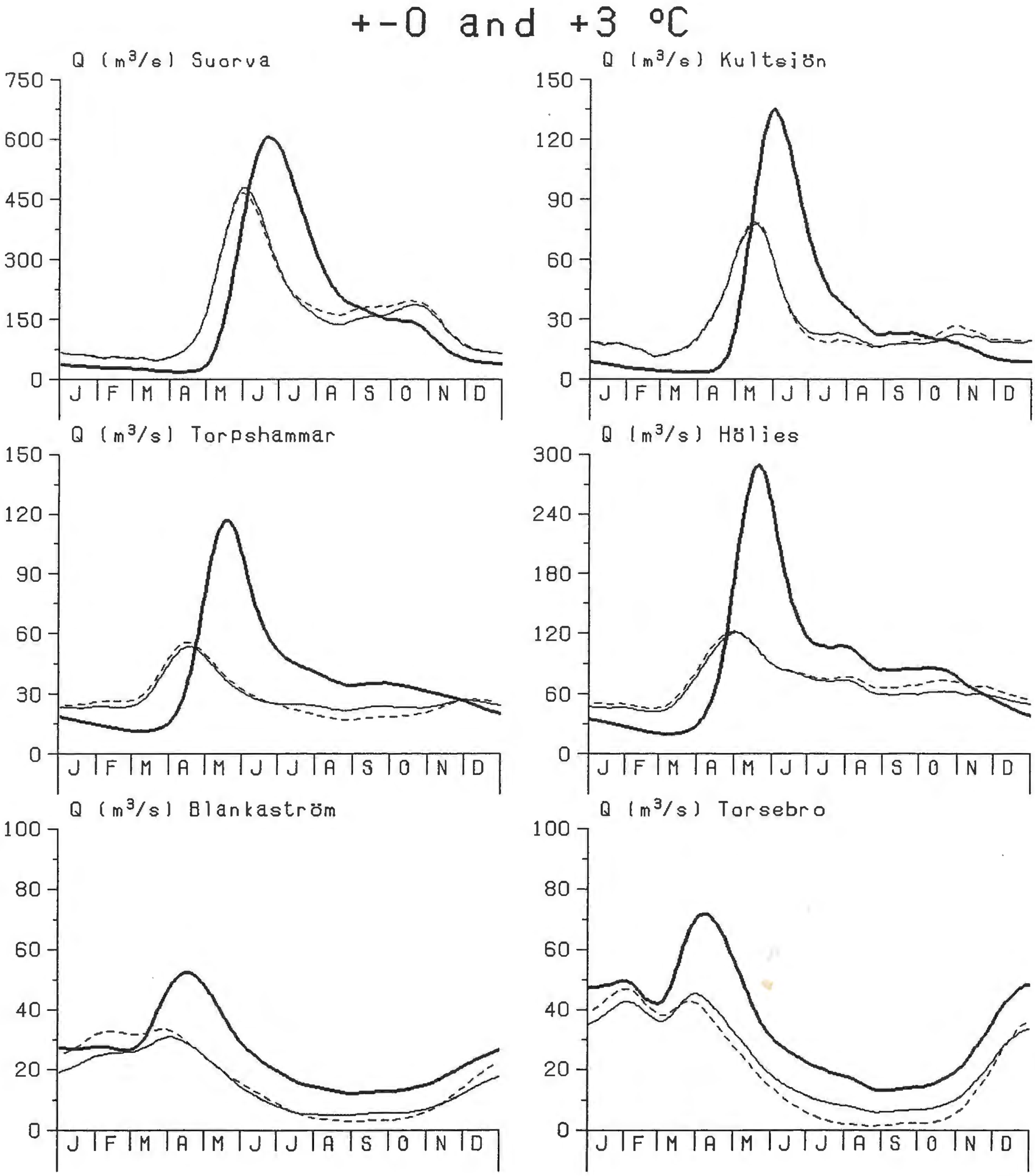

Figure 10. Example of simulated response to a climate change for six of the basins. Daily mean values of computed discharge for the period 1969 -1989, for the present climate (thick line) and after a temperature increase of 3 °C. Solid line = by the F-THO model and dashed line = by the F-ETF model.

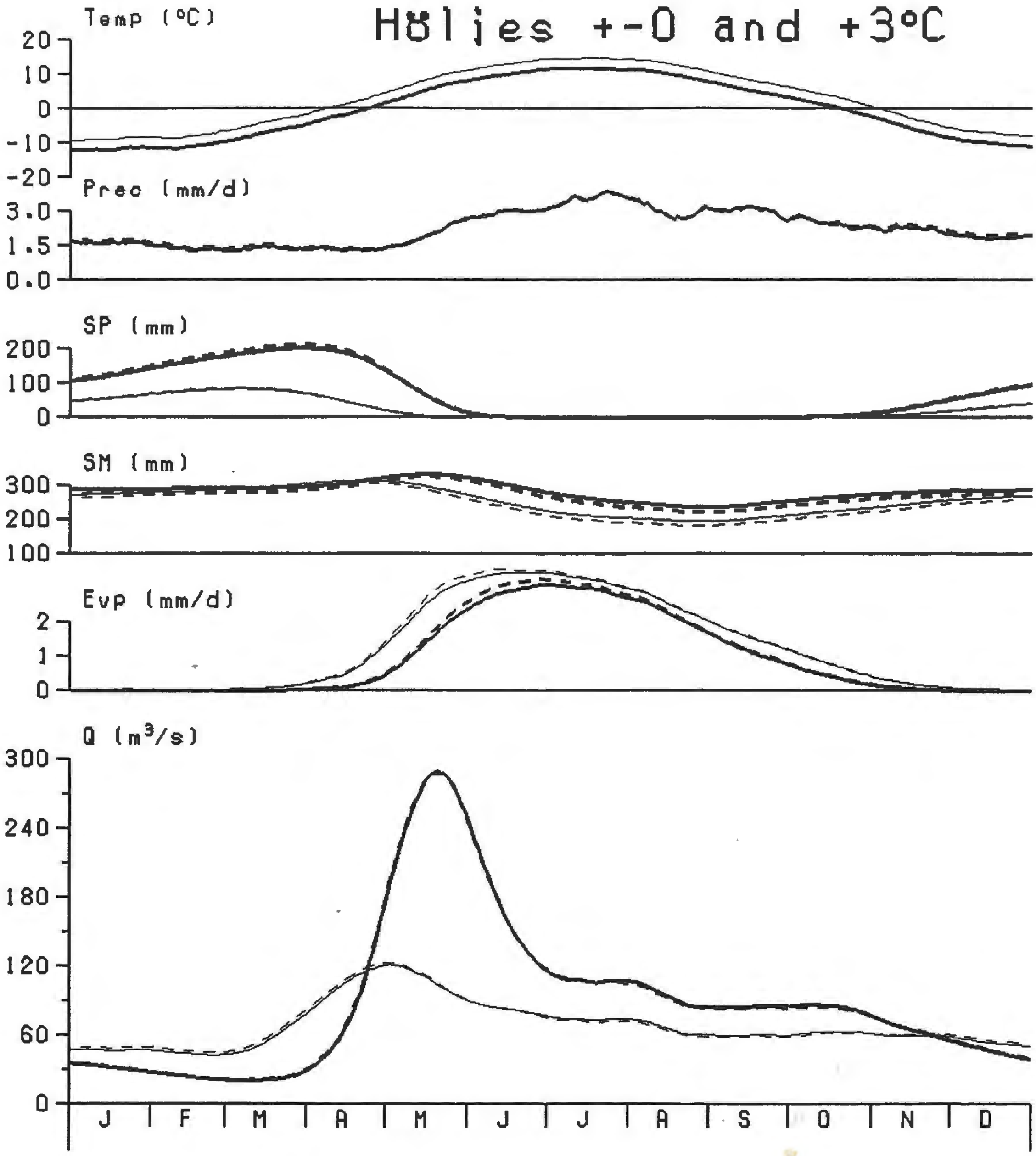

Figure 11. *Additional sensitivity run for Höljes. Average values for the period 1969 - 1989. Thick lines = the present climate according to the standard F-THO model (solid) and the one with increased K_T (dashed). Thin lines = after a temperature increase of 3 °C according to the standard F-THO model (solid) and the one with increased K_T (dashed).*

6. DISCUSSION

Considerably different evapotranspiration routines could be calibrated to yield approximately the same agreement between computed and observed discharge. A more direct comparison was therefore also made between computed and measured potential evapotranspiration. Data from the IHD period were used for this purpose (Figure 12). Both of the two models, ETF and Thornthwaite, were adjusted by a factor until the volume agreed with the measurements, and therefore give different results when compared with the two different types of pans. Considering the difficulty in interpreting pan evaporation as potential evaporation, no reliable conclusion could be drawn from this comparison regarding which model to choose.

The reasonableness of such a simple model as a temperature index (THO) was further investigated. Figures 13 and 14 and Table 6 show comparisons between air temperature and estimates of potential evapotranspiration. The potential evapotranspiration was estimated both with and without using the seasonal varying factor STF. The use of STF gave better agreement in all comparisons, and the results in Table 6 refer to analyses using STF. The resulting values of the parameter K_T and the estimates obtained by model calibration (cf. Table 4) were all fairly near 0.2 mm $°C^{-1}$ day^{-1}. The stability of this parameter is an advantage when it is being used in climate change simulations.

The sensitivity of the model to a change in temperature was considerable. For example, in Blankaström, the simulations indicate that an increase in temperature of 3 °C would lead to a reduction in runoff of some 40 %, a figure which may seem surprisingly high. A rough estimate of the response was therefore also made by studying the water balance map of Sweden (Figure 15). The average temperature difference between the far north and the far south of Sweden is some 10 °C. The Blankaström basin receives about 600 mm of precipitation annually, and the annual evapotranspiration is about 400 mm. In the northern parts of the country, which receive the same amount of rainfall, the evaporation is some 200 mm lower. This gives about 20 mm of increase in evaporation per °C, or 60 mm per 3 °C. A change in runoff by 60 mm amounts to a relative change of 30 % in the Blankaström basin. This number is in the order of the simulated response by the suggested HBV model.

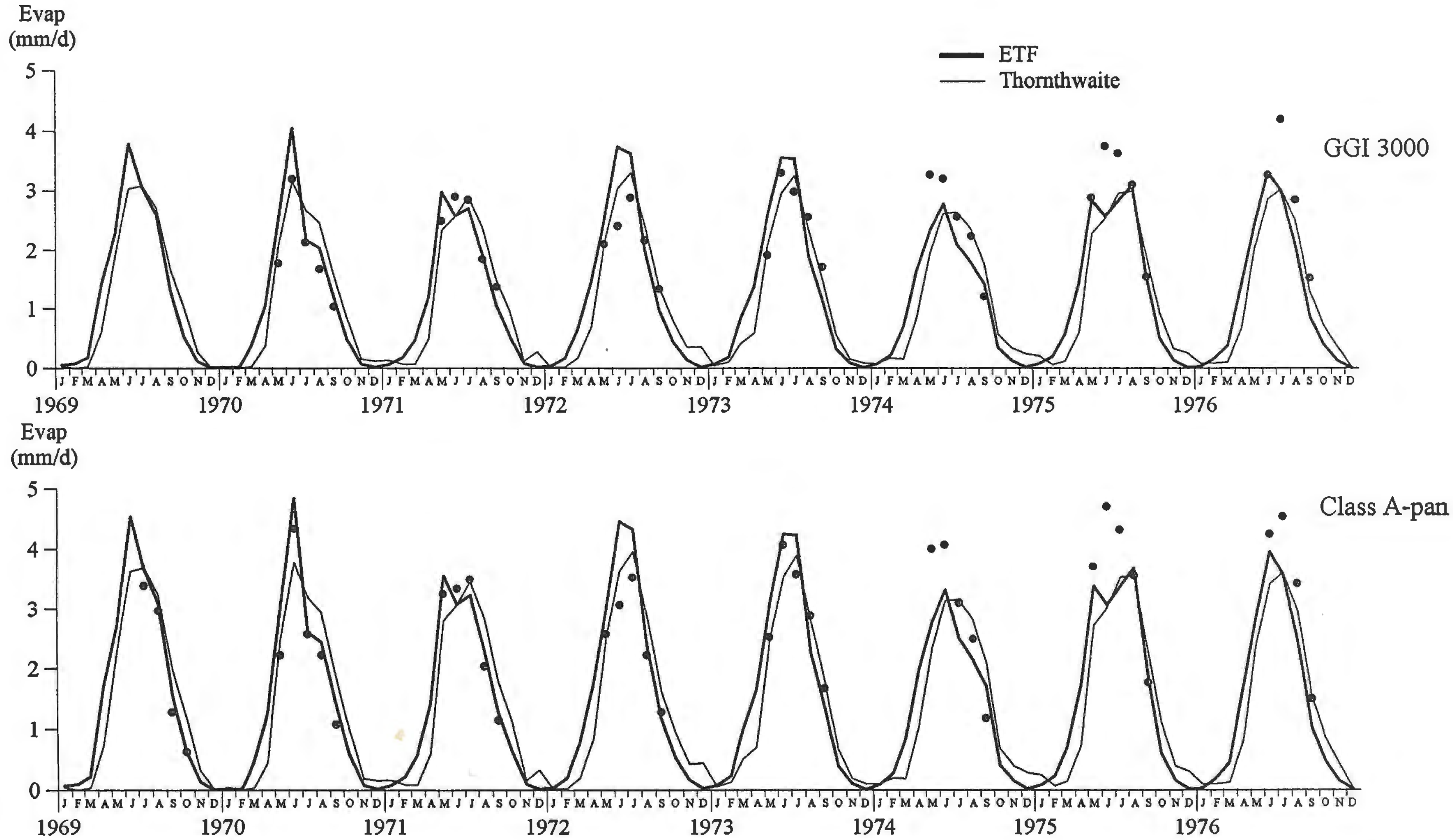

Figure 12. *Computed (by the ETF and Thornthwaite models) and measured potential evapotranspiration (by GGI 3000- and Class A-pans) for the IHD-station Sjöängen in the Velen research area in southern Sweden.*

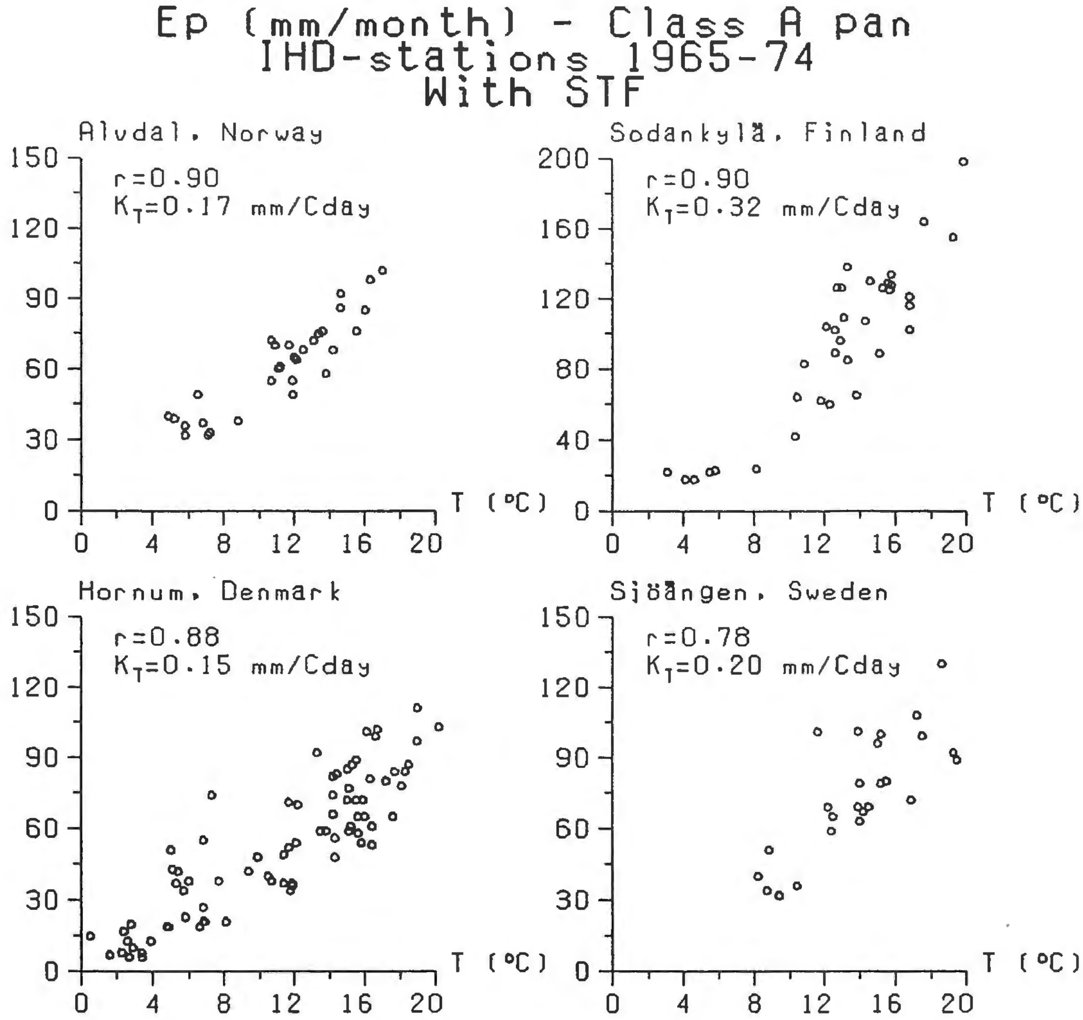

Figure 13. Measured monthly pan evaporation from IHD-stations versus air temperature adjusted by STF.

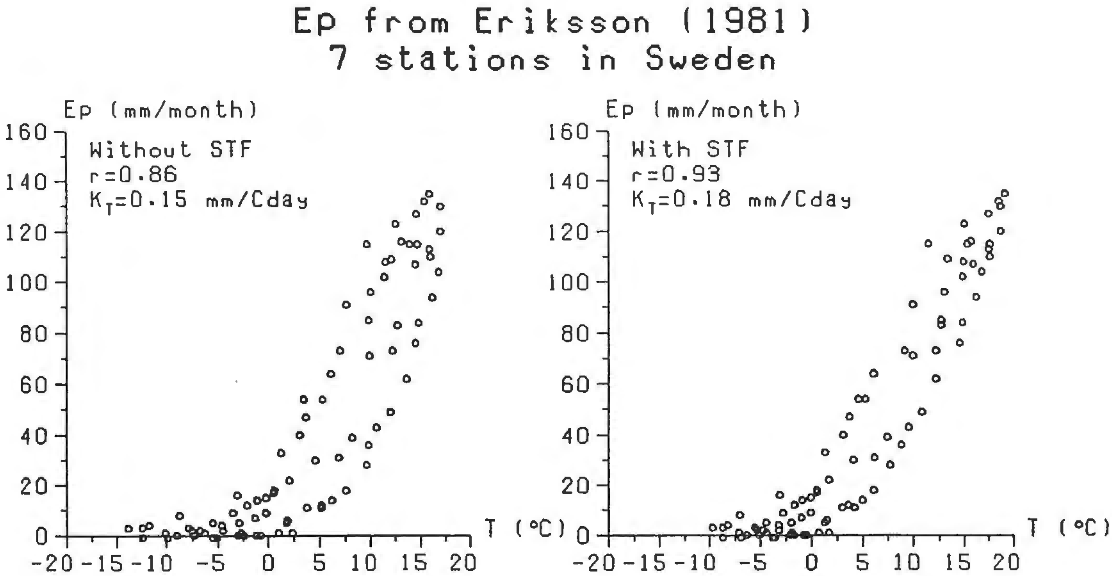

Figure 14. *Computed monthly potential evapotranspiration (Eriksson, 1981) by the Penman formula versus air temperature (without and with adjustment by STF).*

Table 6. *The Thornthwaite parameter K_T (mm °C^{-1} day^{-1}) as obtained from analyses of data from other sources than model calibrations.*

Data source	Basin	K_T
Analysis of data on potential evapotranspiration by the Penman formula by Eriksson (1981)	Kiruna Lycksele Östersund Falun Uppsala Linköping Malmö	0.18
Data from the IHD period	Alvdal Sodankylä Hornum Sjöängen	0.17 0.32 0.15 0.20
Analysis of data on potential evapotranspiration by Bowen ration measurements by Bringfelt (1982)	Velen	0.19

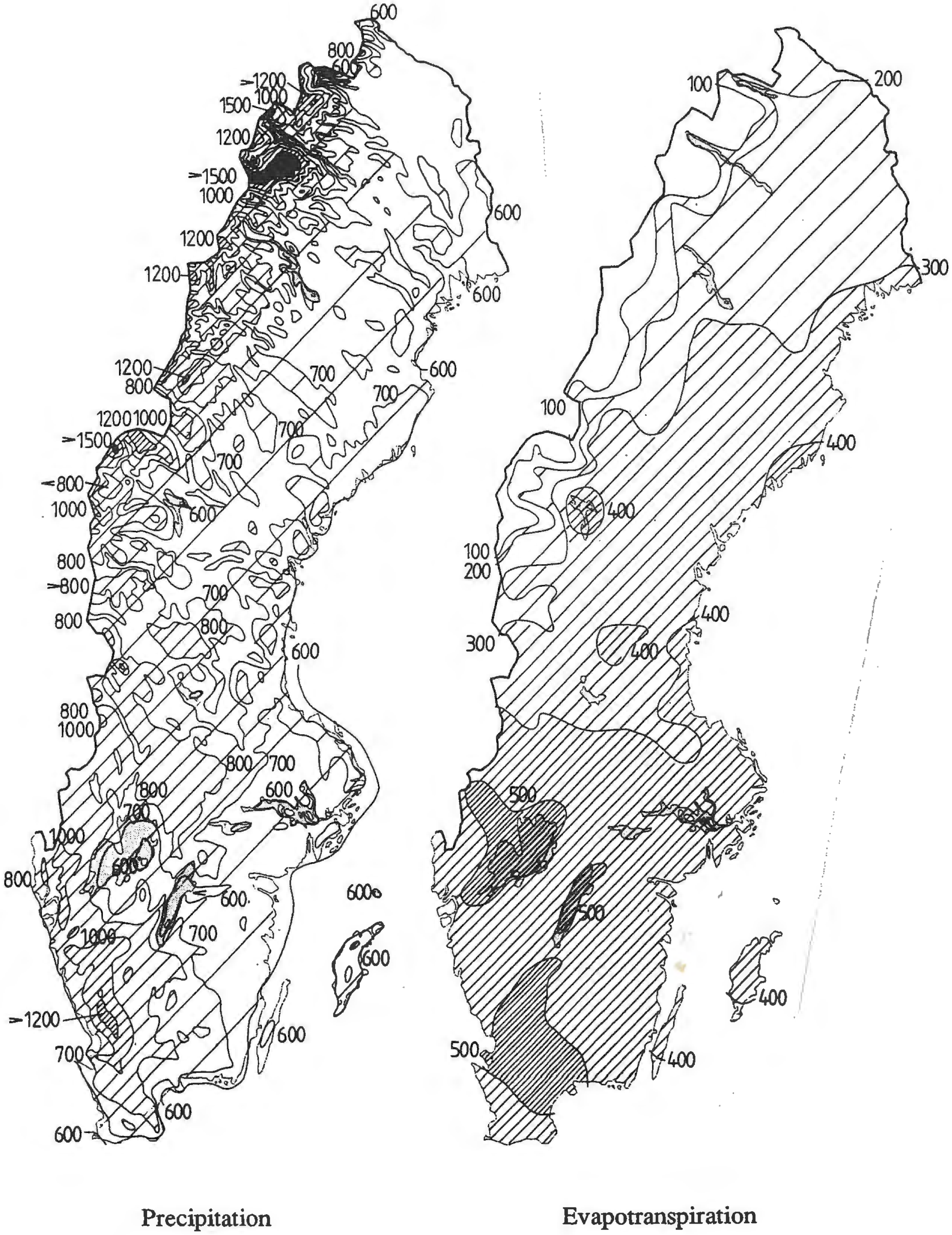

Figure 15. Estimated water balance (in mm/year) for the period 1961 - 1990 (from Brandt et al., 1994).

7. CONCLUSIONS

It was difficult to improve the HBV model. The original model already performed well in most basins, with R^2-values of around 0.8 - 0.9. Andersson (1992) came to the same conclusion when attempting to improve the snow melt and evapotranspiration routines. Some of the changes in the model are realistic and should be kept, even though the simulation results do not seem better. The introduction of more realism does not necessarily improve the simulation results, as old errors may have compensated each other. All models were, however, calibrated. More realistic differences between the evaporation from forests, open areas and lakes should improve the possibilities for computing discharge in ungauged basins, which at present is an important application of the HBV model.

The present interception routine gave poorer results than the original model, but meant that more realistic snowfall correction factors could be used. The unrealistically low correction factors for snow has long been a structural problem in the HBV model.

The automatic calibration technique which was used was a rapid, consistent and objective method for testing and evaluating different model structures. It is, however, difficult to interpret the effects and significance of each model change, and the automatic calibration must be complemented by visual inspection of the results. Some of the automatic calibrations were felt to be very good, whereas the method in some basins failed to remove all systematic errors. The choice of criterion is still a problem with any automatic calibration routine. The same is true for the evaluation and interpretation of the results. The simulation with the highest R^2 was not always felt to be the best one. One specific problem is the handling of the initial state. The snow pack, soil moisture storage etc. in the model depend on the parameters, which are varied in the optimization loop, without any corresponding adjustment of the initial state. This may force the optimization procedure to give preference to some parameter values which agree with the initial state, if too short a period is used. The first period in the calibration period should therefore be excluded from the computation of the criteria.

A simple temperature equation of the Thornthwaite type was nearly as good as the standard monthly Penman estimates. The temperature anomaly routine, with the parameter ETF, gave slightly better results. Its use for climate change simulations is, nevertheless, somewhat more uncertain than the Thornthwaite formula. The parameter in the latter formula is active during the whole calibration period and is better defined by the calibration. The parameter K_T was rather stable, with roughly the same value (0.2 mm °C^{-1} day^{-1}) in all basins and according to some additional comparisons between air temperature and estimates of potential evapotranspiration. The parameter ETF is mostly active during few, extreme weather conditions. Its value is therefore not as well defined by the calibration, but it has a great impact on the response to a temperature change. If ETF is to be used, a regional parameter value should be estimated as was done in this study. Sensitivity analyses, however, indicated that the exact values of model parameters K_T and ETF were not crucial, since all parameters interact in the model. A change in the value of for example K_T is compensated by the changes of the other parameters which are the result of a model calibration.

The Thornthwaite method as used here automatically gives a reasonable reduction in evapotranspiration with altitude. It is simpler than the ETF model. This, together with the greater stability of its parameter, should make the Thornthwaite method more appropriate for simulation of climate change effects. Nevertheless, the two developed models gave fairly similar results in the climate sensitivity analysis. The uncertainty in the predictions of the hydrological response to a climate change, which is due to weaknesses in the hydrological model, is thus probably less important than the uncertainty in the climate change predictions.

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