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Temporal Variation of Evapotranspiration and Growth in Finnish Forest in Relation to Climate

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Abstract

In the boreal zone, the annual amounts of evapotranspiration in forests Ef (mma^{-1}) and the growth of growing stands G (m^{-3}ha^{-1}a^{-1}) are practically determined by climate; the spatial variation of both of them can be largely explained by the effective temperature sum L (°Cd^{-1}) and the maximum soil frost depth in winter F (cm); the latter, in its turn, is a function of snow depth and frost sum. In the southern boreal regions, particularly on sandy glacial tills, drought in early summer is also noticeable. This, which somewhat counterbalances the effect of L, was roughly estimated. The partial derivatives of G with respect to the effective temperature sum L and winter's maximum soil frost depth F were approximated by those obtained earlier by the author from spatial analysis. Applying the relation between Ef and the volume of growing stands K as obtained by Hyvärinen et al. (1995) and the mean ratio G/K on the basis of forest inventories, as well as the derivatives of Ef with respect to L and F, the temporal standard deviations of G and Ef could be roughly approximated for various regions south of the Arctic Circle in Finland. The results were verified by the actual temporal standard deviations of G found in ring-width studies, and evaporation observed by the Class–A pan. The results of this study, albeit rough, were of the right magnitude. It was also shown that the mean snow depth during winter more affected the soil frost depth, Ef and G, than did the frost sum. The correlations between G and the precipitation during winter months, as well as between G and monthly mean temperatures, obtained by ring-width studies, agree well with the influence of snow depth and frost sum obtained in this study.

Key words: variation of forest growth, variation of evapotranspiration, growth of boreal forests, evapotranspiration in boreal forests

1. Introduction

Observations of the annual values of evapotranspiration in forests *Ef* and growth of forest stands *G* are only sporadically available. One way to approach the problem of the temporal variation of *Ef* and *G* is to estimate their temporal standard deviations on the basis of the partial contributions of the main effective climatic variables. In this study, this method was applied to the Finnish mainland south of the Arctic Circle. The starting points used were, on the one hand, the equations derived for the spatial variation of *Ef* and *G*, anchored to values obtained from the water balance equation and national forest inventories, and on the other, temporal variations of the climatic variables involved. The aim of the study was also to reveal the main regional differences arising from different climatic conditions. Consequently, the area studied was divided into six regions. Particularly, the role of soil frost, frost sum and snow depth, often forgotten, were taken into account, in addition to the effects of heat and moisture in summer. Verification values are also needed to ensure that the standard deviations of *Ef* and *G* thus roughly approximated are actually of the correct magnitude. The actual standard deviations are those based on Class–A pan observations and the results of tree-ring studies. The study concentrates on the main features of the effects and variables involved, aiming to achieve preliminary results in order to encourage continued research on the topic.

2. *List of most frequently-used symbols*

ε = the error of estimation, either due to missing variables or the inaccuracy of the method

D = the effect of drought on evapotranspiration in summer, approximated by the difference between unaffected evapotranspiration and precipitation in June or July

Ef = evapotranspiration in the forest stands (mm a^{-1})

E_B = evapotranspiration obtained from the water balance equation of the ground surface

E_6 = evapotranspiration in June (mm $month^{-1}$)

Ef_{basic}= evapotranspiration in a logged area or on pine bogs (mm a^{-1})

E_{ClassA} = evapotranspiration observed by Class-A pan

ΔEf = contribution to Ef by growing forest stands as given by $Ef - Ef_{basic}$ (mm a^{-1})

f = proportion of afforested land area

F = the winter's greatest soil frost depth (cm)

G = growth of forest stands ($m^3ha^{-1}a^{-1}$), without bark

GR = Global radiation (MJm^{-2})

K = volume of growing forest stands (m^3ha^{-1}), with bark

L = the effective temperature sum (°C d)

N = the areal proportion of fens

P = the frost sum ((–°C) d^{-1})

S = mean snow depth during the soil frost period (cm)

T = daily mean temperature over one or several months (°C)

Tw = mean temperature during a winter month (°C)

$Tday$ = day-time mean temperature (approximated here as mean of values at 03, 06, 09, 12, 15, and 18 UTC), averaged over one or several months (°C).

W = the areal proportion of lakes

3. *Method*

3.1 *Growth of tree stands and evapotranspiration in forests as a function of meteorological variables*

The evapotranspiration in forest stands, *Ef*, and growth of boreal forests, *G*, are to a great extent determined by the same climatic factors. In this article, the temporal variation of evapotranspiration in Finnish forests is roughly estimated on the basis of the combined effect of the standard deviations of the climatic factors concerned. The procedure begins by obtaining the values of $\delta G/\delta x_1$ and $\delta Ef/\delta x_1$, where x_1 is the climatic variable considered, from the Eqs. $G = G(x_1,\ldots,x_n)$ and $Ef = Ef(x_1,\ldots,x_n)$ for the areal distribution of *G* and *Ef*. Then, calculating the temporal standard deviation of *x* from the observations, we may approximate the temporal standard deviation of *G* and *Ef* caused by $б\, x_1$ as

$$бG(x_1) \sim б\, x_1 \bullet \delta G/\delta x_1 \tag{1}$$

and

$$бEf(x_1) \sim б\, x_1 \bullet \delta Ef/\delta x_1 \, . \tag{2}$$

When the estimated variables, denoted by $бG(x_1),\ldots,\ бG(x_n)$, and the ignored variables, denoted by $бG(y_1),\ldots,\ бG(y_n)$, are mutually independent, we can write

$$бG\sim ((бG(x_1))^2 +\ldots+ (бG(x_n))^2 + (бG(y_1))^2 +\ldots+ (бG(y_\mathrm{n}))^2+ \varepsilon(x)^2+ \varepsilon^2)^{0.5} \tag{3}$$

and

$$бEf\sim ((бEf(x_1))^2 +\ldots+ (бEf(x_n))^2 + (бEf(y_1))^2 +\ldots+ (бEf(y_n))^2+ \varepsilon(x)^2+ \varepsilon^2)^{0.5} \, . \tag{4}$$

Here, as an approximation, the ignored variables, the random error due to the inaccuracy of the variables x_1 to x_n, denoted by $\varepsilon(x)$, as well the random error connected to the model, denoted by ε, are all assumed to be zero. In cases where there is a significant correlation between variables, such as snow depth and soil frost depth in southern and middle Finland, covariances are also considered. Due to smaller effects not considered here, the estimated values of $бG$ and $бEf$ are rather too low than too high. Note, that in this study, the observed annual values of *G* and *Ef* are, in fact, not explained. However, the estimated values of $бG$ are verified by the observed temporal variation of tree rings and its correlation with climatic variables; similarly, the estimated values of $бEf$ are verified by Class–A pan measurements.

Next, the main variables x_1 to x_n are chosen. Concerning the thermal factors during the growing season, the sum of daily mean temperatures above a certain temperature limit is generally used for this purpose. The limit varies from the freezing point (the sum then being called 'biotemperature') to +5 °C. A better alternative to the traditional temperature sum is, instead of daily means, to consider the day-time temperature sum, i.e., the integration of daily values beginning at a solar elevation of 8 degrees in the morning and ending at a solar elevation of 3 degrees in the evening. The day-time temperature sum takes the global radiation and assimilation better into account than does the traditional measure, and avoids overestimation of the thermal advantage of lake regions due to warm nights, involved in the traditional method (*Solantie*, 2004). Considering that the temporal standard deviations of annual day-time mean temperature during the period May–September are only 5% higher than those for diurnal means, the temporal standard deviations of the traditional and day-time temperature sums are practically the same if the day-time sums are calibrated to the level of the traditional by multiplying by a factor of 1.56. Therefore, the general symbol L is used in this study to denote the effective temperature sum. The winter's greatest soil frost depth, F, is also a factor influencing the growth of forest stands and evapotranspiration through the decomposition and circulation of nutrients, as is soil temperature during the following growing season. The variable F is mainly a function of the frost sum, P, and the winter's mean snow depth, S. One noticeable factor in the southern boreal zone is the drought in early summer, denoted by D. This variable can be expressed as the difference between the unaffected evapotranspiration (i.e., Ef_{basic}, the evaporation over a logged area or a pine bog) and the precipitation in a dry summer month. Instead of July, considered by *Hyvärinen et al*. (1995) for the reduction of evapotranspiration, June was chosen because of its high significance for G. In order to apply Equations (1) to (4), Equations for Ef and G were developed as

$$G = (L, F(P,S), D) \tag{5}$$

and

$$Ef = Ef\,(L, F(P,S), D) = Ef_{basic}(L, F\,(P,S)) + \Delta Ef(G) \tag{6}$$

Here ΔEf is the particular contribution to Ef related to the volume of growing stands and comprising the effect of drought.

3.2 Basic equations for annual evapotranspiration in the forests and growth of forest stands

In Finnish conditions, three main factors are important for the annual evapotranspiration in the forests Ef (mma^{-1}). First, the effective temperature sum is a variable including for a great part, and correlating positively with, such variables as the duration of the vegetation period, the intensity of growth of the biomass, the condensation deficit in the atmosphere and the amount of global radiation (considered in more detail later). It was therefore chosen as the most important variable explaining Ef. Secondly, the volume of growing stands is also important. Considering that the actual evapotranspiration is lower than the potential due to lack of available water, drought must be also taken into account. Using these variables, 93% of the variance of the mean evapotranspiration for Finnish basins over the period 1961–1990, solved from the water balance equation and denoted by E_B, could be explained (*Hyvärinen et al.,* 1995) by the following regression equation

$$E_B = (1 - W) \bullet (a \bullet L + b \bullet f \bullet K + f \bullet D + n \bullet N) + w \bullet W + c\,. \tag{7}$$

Here is W = the areal proportion of lakes, N = the areal proportion of fens, f = the proportion of the afforested land area, a = 0.385, c = –123, $b\,K$ = additional evapotranspiration from growing stands, b = 1.15, and K = the volume of growing stands (m^3ha^{-1}), D = reduction due to drought in early summer, increasing with K and the difference between basic evapotranspiration and precipitation, as well as the occurrence of sand and coarse glacial tills (*Hyvärinen et al*., 1995, *Solantie,* 2004), and n and w are coefficients for fen and lake evapotranspiration.

Considering evapotranspiration in the forests, denoted by Ef, we set $W = 0$ and $N = 0$. With reference to Eq. (6), Ef is a sum of two terms,

$$Ef_{basic} = a\,L + c\,, \tag{8}$$

and

$$\Delta Ef(G) = b\,K + D\,. \tag{9}$$

Regression equation (7) states that the evapotranspiration in forests is given as $Ef_{\mathrm{basic}} + b\,K + D$, that on fens as $Ef_{\mathrm{basic}} + N$ and that on bogs and logged areas as Ef_{basic}. Strictly speaking, the effect of drought (D) does occasionally reduce evapotranspiration in all landtypes; so, application of D to evapotranspiration in forest stands only is an approximation. In summer in a moist boreal climate, water is generally readily available, so that lack of water restricts evapotranspiration appreciably only in that vegetation that needs the most water, i.e., tall forest stands.

Noting that the relative productivity of growing stands in Finland (G), as given by *Ilvessalo* (1960), is approximately

$$G\ (m^3ha^{-1}a^{-1}) \sim 0.038\ K\ (m^3ha^{-1}) \tag{10}$$

we obtain

$$\Delta Ef(G) = 30\ G + D \tag{11}$$

and thus $\delta Ef/\delta G \sim 30$.

On the other hand, *Hyvärinen et al.* (1995), explaining the long-period mean of evapotranspiration E_B, obtained from the water balance equation, by G, using regression analysis, found that $\delta E_B/\delta G \sim 48.4$. This value is, however, an overestimation, because L and K in Equation (7) are spatially positively correlated, as are also L and G.

We may further take into account that the productivity of forests, G, ($m^3ha^{-1}a^{-1}$) in the boreal part of Europe from western Siberia to the Norwegian coast can be expressed as a function of the effective temperature sum L (°Cd), the duration of the vegetation period V (d) and the winter's maximum soil frost depth, F (cm). In Finland, with its small west-east extension, V can be neglected, so that (*Solantie,* 2005)

$$G \sim -1.97 + 0.00647\, L - 0.0435\, F\,. \tag{12}$$

from which $\delta G/\delta L \sim 0.00674$ and $\delta G/\delta F \sim -0.0435$. These values, obtained for the spatial variation, were also applied in this study to the temporal variation; only in southern Finland was $\delta G/\delta L$ somewhat reduced.

As F is a function of the frost sum, P, (– °Cd) and the mean snow depth during the frost season, S, (cm), then G is as well. Solving Equation (10) for K and substituting Eq. (12) in the solution we obtain

$$K \sim -52 + 0.177\, L - 1.14\, F(S, P)\,. \tag{13}$$

This means that the volume of growing stands in forests can be approximately given as a function of climatic variables. The main regional features and magnitude of K obtained from this equation are in a good accordance with those obtained by the national forest inventories (e.g. *Tomppo,* 1998), which shows that climate is more powerful than modern forestry in determining the volume of growing stands in forests. Substituting Eq. (13) in Equation (9), and noting that evapotranspiration in forests is the sum of Ef_{basic} and ΔEf, we obtain the result that Ef can be given as a function of merely climatic variables, i.e., without any knowledge about the volume or growth of growing stands, as

$$Ef\,(\text{mm a}^{-1}) \sim (-123 + 0.385\, L) + D + (-59 + 0.20\, L - 1.30\, F) \tag{14}$$

or

$$Ef\,(\text{mm a}^{-1}) \sim -182 + 0.585\, L - 1.30\, F(S,P) + D\,. \tag{15}$$

For example, in the Eastern Lake-Finland area around Punkaharju with $L = 1200$ °Cd and $F = 10$ cm we find that $Ef \sim 507 + D$ mm. Noting that $D \sim -50$ mm *(Hyvärinen et al.,* 1995) we obtain $Ef \sim 450$–460 mm. In timber–line conditions with $L = 570$ °Cd, $F = 38$ cm and $D = 0$, we obtain, correspondingly, a value of Ef of 150 mm.

The complexity of the interactions between climatic variables, evapotranspiration in the forests and the growth of forest stands is shown in the scheme in Fig. 1.

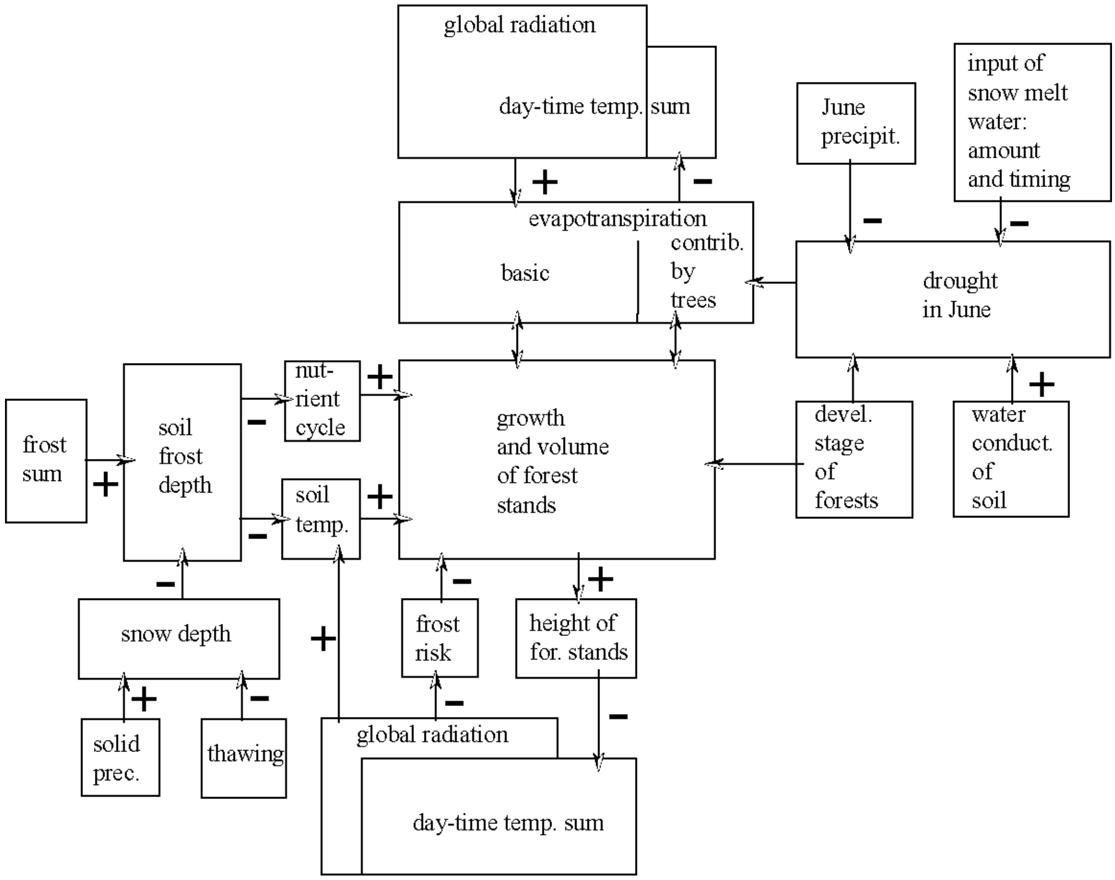

Fig. 1. Showing the complexity of the interactions between climatic variables, evapotranspiration in forests and the growth of forest stands.

3.3 Regions considered

The regions considered were chosen on the basis of both summer and winter conditions. Considering summer, southern boreal areas (SB), where drought is effective, were separated from the middle boreal ones (MB), where drought is negligible. Considering winter conditions, both the increase of P and the decrease of $\delta F/\delta P$ north-eastwards, as well as the increase of S and decrease of $\delta F/\delta S$ north-eastwards, were taken into account. On this basis, the area considered was divided (Fig. 2) into the west coast, the Uusimaa-Häme region, Eastern Lake-Finland (these three in the SB), Eastern Bothnia, the Suomenselkä region, and the Kainuu region (the latter in the MB).

In the Kainuu region, the mean snow depth on January 15th is 45 cm or more (*Solantie,* 2000). In the Suomenselkä region and Eastern Lake-Finland it is 35 to 44 cm, in Eastern Bothnia and in the Uusimaa-Häme regions 25 to 34 cm, and on the west coast 15 to 24 cm. In the Kainuu region, the correlation coefficient $r(S,P)$~–0.2, which means that the frost sum and snow depth are practically independent of each other, while in the other regions, $r(S,P)$ is 0.5 to 0.7 (*Solantie and Drebs,* 2001).

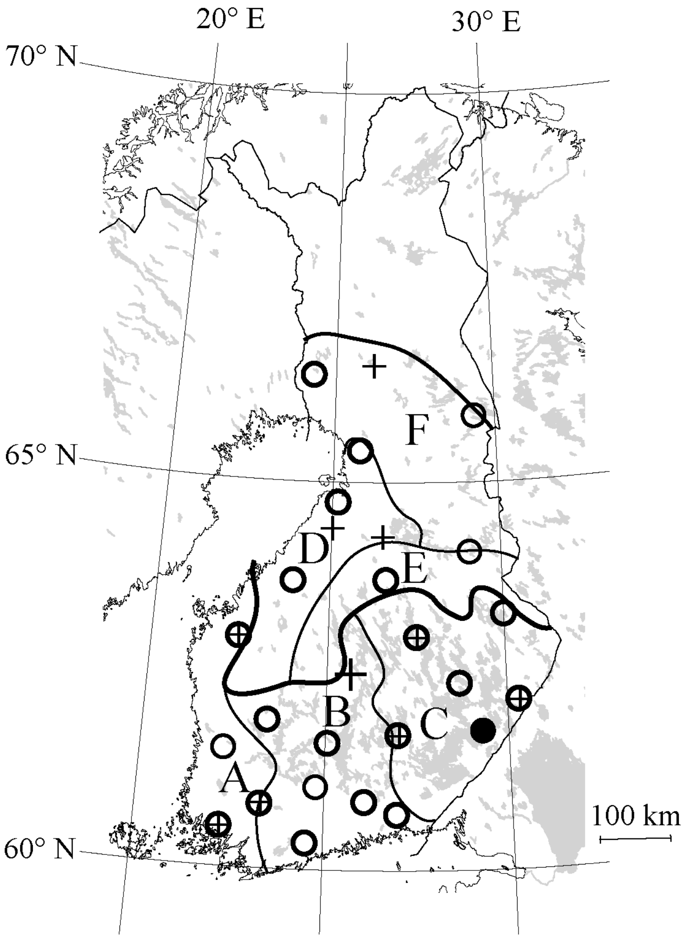

Fig. 2. The locations of stations for evaporation measurements by Class-A pan (crosses, large cross for Luonetjärvi) and for soil frost observations (bold open circles: accepted, thin circles: rejected), and the Punkaharju site with tree-ring studies by *Helama et al.* (2006) (filled circle). Solid lines envelop regions whose variables are used to approximate the annual temporal variations in forest growth and evapotranspiration. The thick line denotes the boundary between the southern boreal zone, where drought in early summer is effective, and the middle boreal zone, where it is not.
A = West coast, B = Häme-Uusimaa region, C = Eastern Lake-Finland, D = Eastern Bothnia, E = Suomenselkä region, and F = Kainuu region

3.4 *Temporal standard deviation of the soil frost depth*

The role of *S* and *P* in the variation of *G* is appreciably higher than that of *L*.

The effect of the winter's mean snow depth *S* (cm) on the temporal variation of the winter's greatest soil frost depth *F* (cm) can be given as

$$\sigma F(S)p = \sigma S \bullet \delta F/\delta Sp \qquad (16)$$

and the corresponding effect of the frost sum *P* (–°C) as

$$\sigma F(P)s = \sigma P \bullet \delta F/\delta Ps\ . \qquad (17)$$

In Equations (16), (17) and (18), σF, σS and σP are the temporal standard deviations of *F*, *S* and *P*. In the partial derivative of Eq. (16), *P* is kept constant while in that of Equation (17), S is kept constant, as indicated by subscripts. Removing the subscripts for brevity, we obtain

$$бF = ((\delta F/\delta S) \bullet бS)^2 + (\delta F/\delta P) \bullet бP)^2 + 2\ \delta F/\delta S\ \delta F/\delta P \text{ cov } (P,S))^{0.5}\ . \tag{18}$$

where the first and second terms (variances) are appreciably larger than the third (covariance) term.

Considering the covariance, we may note that in the southern and central parts of Finland, up to the south-western boundary of the Kainuu region (Fig. 2), thawing in winters milder than average appreciably reduces snow depth: There, the insulation provided by the snow cover increases with increasing frost sum, which reduces the temporal variation of soil frost. On the other hand, in the Kainuu region, snow depth is slightly positively correlated with the winter's mean temperature, because an increase in temperature enhances the amount of solid precipitation more than snow melting gnaws away at the snow cover.

The values of $\delta F/\delta Sp$ and $\delta F/\delta Ps$ were obtained changing values of S respectively P around their long period means, keeping P respectively S constant, and obtaining the corresponding changes in F from the equation by *Solantie* (2005), being a modification for Finnish forests of the soil frost equation derived by *Andersson* (1964):

$$\mathrm{F} = (40\ P + (1.4 \ln P \bullet \mathrm{S})^2)^{0.5} - 1.4 \ln P \bullet S - 0\ 03\ L\ , \tag{19}$$

The values of the coefficients in equation (19) had been adjusted by Solantie using the observed 20-year means of *Huttunen and Soveri* (1993), so as to correspond to the spatial variation of these means. The values of S for the calculation had been those obtained from the soil frost observation sites as given by *Huttunen and Soveri* (1993), while the values of P had been obtained by interpolating between values at climatic stations. In order to ensure that Equation (19) works properly in respect of temporal variations in F, values of бF, based on the observed annual values of F for the period 1971–1990 (*Huttunen and Soveri,* 1993), were compared with the corresponding values obtained by Equation (19) at the same sites.

Concerning the observed values of бF for this comparison, four of the nineteen available series were observed in soils of sandy moraine (Lammi, Kokemäki, Kuhmo and Kuusamo). These stations were excluded because F and бF in such soils are greater than in the other soils.

Finally, Equation (18) was applied to the climatological observation material of S and P to obtain regional means of бF for the period 1961–2000. As basic material for S, annual grid point values for the period 1963–1998 (*Solantie,* 2000) were used; S was approximated by the mean snow depth on January 15th + 2 cm. The spatial variation of бS reflects the orographical variation in precipitation. It is greatest in the Uusimaa-Häme region where orographical enhancement in precipitation is accentuated during south-easterly winds; during such winds, thawing is less than during south-westerly winds (*Solantie and Drebs,* 2001). The temporal standard deviation of the mean snow depth on 15th January is very close to the temporal standard deviation of the mean over the whole winter. For example, at the Punkaharju station during the period 1964–1998, the temporal standard deviation on 15th January was 14.3 cm, while that of the mean of the depths on the 15th of December, January, February, and March was 13.5 cm. The values of $\delta F/\delta S$ and $\delta F/\delta P$ were obtained using Equation (19) for each value of S and P considered. Generally speaking, the values of $\delta F/\delta S$ and $\delta F/\delta P$ decrease with increasing values of S and P. The values of σP are about equal all over the country; the value 320 –°Cd, obtained as an average for the winters of 1946/47–1992/93 studied by *Solantie and Drebs* (2001), was applied in all cases. At Punkaharju, a value of 323 –°Cd was obtained for the period 1964–1998.

3.5 Estimation of the temporal variation of the annual growth of forest stands

The standard deviation of the growth in forests, σG, can be given as

$$\sigma G \sim ((\sigma G\ (L,D))^2 + \sigma G\ (F))^2)^{0.5} \tag{20}$$

where, referring to Eq. (12),

$$\sigma G(F) \sim -\delta G/\delta F \bullet \sigma F \sim 0.435\ \sigma F\ . \tag{21}$$

Further, we have

$$\sigma G(L,D) \sim ((\sigma G(L)^2 + (\sigma G(D)^2)^{0.5}\ . \tag{22}$$

Considering that drought (D) is insignificant in the middle boreal zone with a moist climate, and referring to Equation (12) for $\delta G/\delta L$, we obtain

$$\sigma G\ (L,D) \sim \sigma G\ (L) \sim \delta G/\delta L \bullet \sigma L \sim 0.00647 \bullet 130 = 0.84\ \mathrm{m^3ha^{-1}a^{-1}}\ . \tag{23}$$

In the southern boreal zone, the correlation between temperature and growth of forest stands in early summer is highly negative; for the temporal variation it can be roughly estimated that $\delta G/\delta L$ is roughly a quarter of that in the middle boreal zone. Consequently, $\sigma G \sim 0.2\ \mathrm{m^3ha^{-1}a^{-1}}$.

To estimate the effect of drought (D) on σG, we begin with the standard deviation of precipitation in June as $\sigma R_6 \sim 25$ mm and the correlation coefficient $r(G, R_6) \sim +0.50$ (*Helama et al*., 2006). Assuming that a change in R_6 causes an equal but opposite change in June evapotranspiration E_6, so that $\sigma E_6 \sim \sigma R_6 \sim 25$ mm, we have $r(G, E_6) \sim -0.50$. Considering further that $\delta E/\delta G \sim 30\ \mathrm{mm/m^3ha^{-1}}$ (Eq. 11), we obtain the result that $\sigma G\ (E_6) \sim 0.5\ (\sigma E_6/(\ \delta E/\delta G)) = 0.38\ \mathrm{m^3ha^{-1}a^{-1}}$. Consequently, in the southern boreal zone, $\sigma G(L,D) \sim (0.2^2 + 0.38^2)^{0.5} \sim 0.43\ \mathrm{m^3ha^{-1}a^{-1}}$.

3.6 Estimation of the temporal variation of the annual evapotranspiration in forests

The standard deviation of evapotranspiration in forests, $бEf$, can be given as

$$бEf \sim ((бF \bullet \delta Ef/\delta F)^2 + (бL \bullet \delta Ef/\delta L)^2 + бEf(D))^{0.5} . \tag{24}$$

where

$$\delta Ef/ \delta F = \delta Ef/\delta G \bullet \delta G/ \Delta F . \tag{25}$$

Referring to Eq. (6) we have

$$\delta Ef/ \delta L = \delta Ef_{basic}/\delta L + \delta\Delta Ef/\delta \mathrm{L} = \delta Ef_{basic}/\delta L + \delta Ef/\delta G \bullet \delta G/\delta L . \tag{26}$$

Substituting the values of the variables involved in Equations (24) to (26) we obtain approximations for $бEf$.

4. Results

4.1 Comparison between observed and calculated soil frost depths

Comparison with observed and calculated values of $бF$, excluding those at sites where the soil is sandy glacial till, gives equal results:

Mean ± standard deviation of the observed values of $бF$: 18.5 ± 7.6

Mean ± standard deviation of the calculated values of $бF$: 18.5 ± 5.8.

The regression equation between the observation-based and calculated values of $бF$ is

$$бF_O = -1.1 + 1.1\ бF_C \tag{27}$$

with a correlation coefficient of 0.81.

In the Kainuu region, the observed values at the Kuhmo and Kuusamo sites on sandy glacial till of 19.2 and 17.0 cm are somewhat higher than the respective calculated values of 15.0 and 13.5 cm, while at Ylitornio, where the soil is fine sand, the observed value of $бF$, 17.5 cm, is close to the calculated 15.6.

4.2 Temporal standard deviation soil frost depth by regions

Using climatic snow depth statistics for S (*Solantie*, 2000) to calculate the values by regions of the variables in Equation (18), we obtain the temporal standard deviations of soil frost given in Table 1.

Table 1. Calculation of the temporal standard deviation of the winter's greatest soil frost depth σF as a function of variables involved in Eq. (18) by regions (Fig. 2). Here S denotes snow depth, P frost sum and σS and σP their temporal standard deviations.

Region	cov (S,P)	S cm	P °Cd	σS cm	σP °Cd	$\delta F/\delta S$	$\delta F/\delta P$	σF cm
West coast	1200	20	500	13	320	–1.85	0.074	28.5
Uusimaa-Häme	1700	27	620	16	320	–1.48	0.060	25
Eastern Lake–F.	1000	40	920	14	320	–0.95	0.037	15.5
Eastern Bothnia	750	30	820	13	320	–1.46	0.049	22.5
Suomenselkä	500	40	920	14.5	320	–0.95	0.037	17
Kainuu region	–450	53	1360	15	320	–0.80	0.029	16

We note that the values of σF decrease from the south-west towards the north-east due to the common effect of the spatial distributions of $\delta F/\delta P$, $\delta F/\delta S$ and cov (S, P). The absolute values of $\delta F/\delta P$ decrease towards the east-north-east with increasing snow depth. The values of cov (S, P) decrease broadly northwards due to the common effect of thawing and the correlation between precipitation and wind direction caused by orography. Thawing, enhancing the covariance, increases towards the south-west. In some regions, particularly in Uusimaa-Häme and somewhat also in Eastern Lake-Finland, precipitation falls more during south-easterly winds during which thawing is less frequent than during south-westerly winds, while in Eastern Bothnia, and particularly in the Kainuu region, the opposite is true (*Solantie and Drebs*, 2001). The correlation coefficient between snow depth and frost sum is highest in the county of Uusimaa (*Solantie and Drebs*, 2001), as too is the value of cov (S,P). The effect of the term $\delta F/\delta P \bullet \delta F/\delta S \bullet$ cov (S, P) in Equation (18) on σF is greatest, –5 cm, in Uusimaa, and also on the west coast, where the absolute values of $\delta F/\delta P$ and $\delta F/\delta S$ are highest due to the thin snow cover.

4.3 Temporal standard deviation of forest growth by regions

Substituting the values of σF in Table 1 into Equation (21), we obtain the result that $\sigma G(F)$ varies from 0.7 $m^3ha^{-1}a^{-1}$ in Eastern Lake-Finland, Suomenselkä and the Kainuu region to 1.2 $m^3ha^{-1}a^{-1}$ on the west coast. These are appreciably higher than the values of σG (L,D) of 0.43 $m^3ha^{-1}a^{-1}$ in the southern boreal regions, but of the same order as σG (L,D) of 0.84 $m^3ha^{-1}a^{-1}$ in the middle boreal regions.

The values of σG and the proportions of its various components (% of σG) are given in Table 2.

In conclusion, that part of the standard deviation of the forest growth, explained by the annual variation of the considered climatic variables, seems to be within the limits 1.05 ± 0.25 $m^3ha\ ^1a^{-1}$. It seems to be smallest in Eastern Lake-Finland and greatest on the west coast and in Eastern Bothnia.

Table 2. The estimated standard deviation of forest growth σ*G* ($m^3ha^{1}a^{-1}$) by regions (Fig. 2), based on partial standard deviations, as a function of frost sum *P*, snow depth *S*, drought *D* and the effective temperature sum *L*. Each partial standard deviation is given as a percentage of the variance explained by it.

Region	σ*G* $m^3ha^{1}a^{-1}$	σ*G*(*P*)/σ*G* (%)	σ*G*(*S*)/σ*G* (%)	σ*G*(*D*)/σ*G* (%)	σ*G*(*L*)/σ*G* (%)
West coast	1.31	45	45	8	2
Uusimaa-Häme region	1.17	34	53	10	3
Eastern Lake-Finland	0.80	29	42	23	6
Eastern Bothnia	1.29	24	34	0	46
Suomenselkä+Kainuu	1.11	16	26	0	58

The most important factors explaining the temporal variation of the growth of forest stands seem to be on the west coast: snow depth and frost sum, in the Uusimaa-Häme region and in Eastern Lake-Finland: snow depth, and in the Kainuu and Suomenselkä regions and in Eastern Bothnia: the day-time temperature sum.

On the other hand, considering the importance of the elements separately, the frost sum is relatively most important on the west coast with a 45% share, the snow depth is relatively most important in the Uusimaa-Häme region with a 53% share, drought is relatively most important in Eastern Lake-Finland with a 23% share and the effective temperature sum is relatively most important in the Kainuu and Suomenselkä regions with a 58% share.

On the west coast, the snow depth affects soil frost and the growth of forests as much as the frost sum but in the other regions appreciably more.

Note that the time series of *P* in the southern boreal zone, L in the middle boreal zone and *S* in both of them could all be reconstructed with the same accuracy as was the time series for June precipitation at Punkaharju, constructed by *Helama et al*. (2006) on the basis of tree-ring widths.

4.4 Comparison of the results in this study with studies on ring-width chronology

Ring-width chronology bears evidence of the role various climatic factors play in the temporal variation of the growth of forest stands. *Helama and Lindholm* (2003), and *Helama et al.* (2006) have calculated correlations between the tree rings of Scots pine (*Pinus sylvestris* L.), on the one hand, and monthly mean temperatures and monthly amounts of precipitation, on the other, in the region around Punkaharju (61.8 °N, 29.3 °E) in Eastern Lake-Finland. They found, in accordance with the results in the previous section, that climatic conditions in winter and drought in early summer, and not only the effective temperature sum, are significant for the growth of trees, and thus also for evapotranspiration. *Mielikäinen* (1996) has shown that drought in June also plays a crucial role in the growth of Southern-Finnish spruce.

Helama et al. (2006) found that the correlation between the annual ring-width, on the one hand, and the monthly mean temperatures from December–April, on the other, was on average +0.22. Noting further that the correlations between the annual width of tree rings and the amounts of precipitation in October, December, February and March were of the same magnitude, the connection between the annual growth of trees and the winter's greatest soil frost depth is obvious. On the other hand, considering that the frost sum, among the variables considered, explains 28% of the total variance of G in Eastern Lake-Finland, and that the correlation between P and the mean temperature of any winter month, denoted by Tw, is about -0.67, we find that the upper limit for $r(Tw,G)$ is $0.67 \bullet 0.28^{0.5} = +0.35$.

The correlations between the width of tree rings and August and September mean temperatures were also of this magnitude, while the correlation with the July mean temperature was zero and with the June mean temperature was even negative ($r = -0.19$) which illustrate the effect of drought in early summer. Firmer evidence still of the effect of drought on the growth of trees is provided by the highly positive correlations between the tree ring width and the precipitation amounts in May (+0.32) and June (+0.50). On the basis of the latter relationship Helama et al. even roughly reconstructed estimations of June precipitation backwards in time to before the period of climatic observations. Note, however, that at Punkaharju the effect of drought is accentuated both due to the soils and the climate.

Mielikäinen (1996) found the coefficient of variation of annual growth to be 23% for 90–120 year-old pines during the period 1893–1993 in Central-Finnish nature reservations, while for Southern-Finnish spruce stands the relative variation was 10%. In the southern and middle boreal zone on average, the temporal standard deviation of the annual growth of forest stands, without bark, calculated as function of the climatic variables considered, is 1.10 $m^3ha^{-1}a^{-1}$ as a weighed average of the values in Table 2, or about 29% of the mean growth at present in the forests of the southern and middle boreal areas (*Finnish Forest Research Institute,* 2004), the latter being 3.8 $m^3ha^{-1}a^{-1}$, reduced by 21% from the figure of that with bark. Because the responses to the effective climatic variables are distributed over several years (e.g. *Mielikäinen,* 1996, *Lindholm et al.*, 2000), the actual standard deviation and coefficient of variation of annual growth are somewhat lower than those estimated in this study.

Concerning the regional variation of forest growth, *Lindholm et al.* (2000) observed ring-widths of Scots pine during the period 1806–1991 from samples situated 1) at the timber-line, 2) at the southern edge of the middle boreal zone both in east and west and 3) in the Punkaharju region within the southern boreal zone, but none in the southern and western regions with thin snow cover. Their conclusion that 'Going from the north to the south, the variance of ring-width series becomes smaller' is in accordance with the fact that the variance observed in this study in the Kainuu region (between the two northernmost and middle sites of Lindholm et al.) was clearly greater than that in the Punkaharju region in Eastern Lake-Finland. Also the observation by *Lindholm et al.* (2000) that 'growing season temperatures govern the growth rates of northern pines, while towards south, pine growth becomes less affected by temperatures, and more affected by e.g. precipitation' agrees well with the results in this study. Further, *Lindholm et al.* (2000) found at all groups of sites that the correlations between monthly mean temperatures from November to March and ring-width were positive in all regions of Finland, which is firm evidence of the significance of the frost sum for the growth of forest stands. Helama et al. found as well that ‘Unlike the north and south, the middle zone’ (sites 2 above at the southern edge of the middle boreal zone) ‘did not yield any clear and unambiguous signals. It is possible that our samples from the middle zone contain a mixture of several signals, none of which dominates. Some of the signals may have their origin in other non climatic, stand wise sources’. This observation can be explained by the results of this article. First, at the southern edge of the middle boreal zone the effective temperature sum is less effective than farther north. Secondly, the effect of frost sum and snow depth on the growth somewhat counterbalance each other’s effect on the growth, and thirdly, the effect of drought exists, but weaker than in the Punkaharju region, so that none of the effects is dominant. In all regions, prior growth, mostly one year before but to some degree also 2 and 3 years earlier, was significant for the growth during the year considered. In other words, the growth of forest stands during any particular year is influenced by the climatic factors during the preceding 1–3 years.

4.5 Temporal standard deviation of the annual evapotranspiration in forests

Substituting the values given below of the variables involved in Equations (24) to (26), we may obtain the approximations for бEf by regions:

$\delta Ef/\delta G$ = 30 mm/m^{3}ha^{-1}a^{-1}, $\delta G/\delta F$ ~ –0.0435 m^{3}ha^{-1}a^{-1}cm^{-1}, $\delta Ef_{basic}/\delta L$ = 0.385 mm/°Cd, in the middle boreal regions $\delta G/\delta L$ (m^{3}ha^{-1}a^{-1}/°Cd^{-1}) = 0.00647, while in the southern boreal regions its value is 0.0015, бL ~ 130 °Cd, б$Ef\,(D)$ ~ 25 mm. Considering winter conditions, $\delta G/\delta F$ ~ –0.0435 m^{3}ha^{-1}a^{-1}cm^{-1}.

The standard deviation of evapotranspiration without any reduction due to drought in tall tree stands or the effect of winter conditions can by approximated by б$Ef_{basic}(\delta L)$ ~ 0.385 • 130 = 50 mm. This standard deviation of Ef is of the same magnitude as that observed in 1961–1990 by Class–A pans at the observation stations then in use, i.e. Vaala, Ruukki, Rovaniemen mlk., Mikkeli, Tohmajärvi, Maaninka; Mietoinen, Jokioinen and Ylistaro (Fig. 1): on average 50.6 mm (*Järvinen and Kuusisto,* 1995).

Concerning the particular contribution by forest stands, we refer to Equations (11) and (12). In the middle boreal regions, $\delta\Delta Ef/\ \delta L$ • бL ~ $\delta\Delta Ef/\ \delta G$ • $\delta G/\delta L$ • бL = 30 • 0.00647 • 130 = 25.2 mm, while in the southern boreal regions the corresponding value is about 30 • 0.0015 • 130 = 6.3 mm; we now have the standard deviations of Ef caused by F, L, and D and the proportions (%) of their common contribution to the total standard deviation of Ef, as given in Table 3:

Table 3. The numerical and percentual contributions of F, L, and D to the total standard deviation of Ef.

	$\delta Ef/\delta F \cdot \sigma F$		$\delta Ef/\delta L \cdot \sigma L$		$\delta Ef/\delta D \cdot \sigma D$		σEf
Kainuu region	21.0	7%	75.2	93%	0		78.1
Suomenselkä	22.0	8%	75.2	92%	0		78.4
Eastern Bothnia	29.5	13%	75.2	87%	0		80.8
Eastern Lake-Finland	20.0	10%	56.0	75%	25.0	15%	64.5
Uusimaa-Häme	32.5	22%	56.0	65%	25.0	13%	69.4
West coast	37.0	27%	56.0	61%	25.0	12%	71.6
Average	27.0	14%	65.6	79%	12.5	7%	73.8

Taking into account that the behaviour of climatic variables in any individual year influences ΔEf for the following few years, we may conclude that these values are overestimates of the annual variation. Assuming mutually-independent yearly values, the standard deviations of 30-year means of Ef would be 12 to 15 mm, actually somewhat smaller.

For the temporal variance of the evapotranspiration in forests, the role of the effective temperature sum is overwhelming in the middle boreal zone, also explaining ⅔ to ¾ of the variance in the southern boreal areas. The effect of winter conditions is highest in the southern and western regions, while the effect of drought is highest in Eastern Lake-Finland.

Let us further consider how closely the effective temperature sum reflects the temporal variation of evapotranspiration. As we just found, the temporal standard deviation of Ef_{basic} (calculated as a function of the effective temperature sum L) is about equal to that of Class-A pan evapotranspiration, denoted by E_{ClassA}. Additionally, it is also necessary to study the correlations between Ef_{basic} and E_{ClassA}, and the correlations between each of the evaporation variables and the global radiation GR. As the values of E_{ClassA} and GR are given as monthly values, and since $L = T\ V$, where V = the duration of the vegetation period and T = the mean temperature during this period in °C, we may, in this connection, replace L with T. In order to compare results obtained by the traditional temperature sum, obtained as a function of T, and the day-time temperature sum, we also calculated the correlations when using a mean temperature found by averaging over day-time observations, i.e., at 03, 06, 09, 12, 15 and 18 UTC (1.6 hrs behind the solar time), denoted by $Tday$. Consequently, we consider the correlations between the variables involved for the five-month period May to September, as well as for the month of June, as then the solar radiation and evapotranspiration are at their highest. Using observations at Luonetjärvi (Jyväskylä airport), situated in the middle of the stations and regions studied (Fig. 2), for the period 1971–1990 both for E_{ClassA} (*Järvinen and Kuusisto,* 1995) and GR (*Finnish Meteorological Institute,* 1982, 1993) as well as for T and $Tday$, we obtain correlation coefficients as follows:

	$r(Tday, E_{ClassA})$	$r(T, E_{ClassA})$	$r(GR, E_{ClassA})$	$r(GR, Tday)$	$r(GR, T)$
June	0.70	0.67	0.91	0.64	0.60
May–Sep.	0.44	0.41	0.87	0.52	0.45

We may note that all of the correlations are positive, and are higher for June, the month of the highest evapotranspiration, than for the five–month period. We also note that the correlations for *Tday* are higher than for *T*, both with E_{ClassA} and *GR*. The variable *Tday* explains half of the variance of E_{ClassA} and 40% of the variance of *GB*: Consequently, the day-time temperature sum is more skilled in obtaining the annual variation of the actual evapotranspiration than is the traditional sum. Note also the high correlations between E_{ClassA} and *GR*, being obviously higher than the correlations between the actual evapotranspiration in the forests and *GR*, respectively. On the other hand, the correlations of *Tday* with E_{ClassA} are greater than with *GR*, both in June and during the five–month period

4.6 Examples of the effect of the changes in the growth of forest stands and evapotranspiration in forests as a response to changes in snow depth and frost sum

In this section, examples are given of the effect of snow depth and frost sum on evapotranspiration and the growth of forest stands.

Substituting in Equations

$$\delta G/\delta S \sim \delta F/\delta S \bullet \delta G/\delta F \tag{28}$$

and

$$\delta G/\delta P \sim \delta F/\delta P \bullet \delta G/\delta F \tag{29}$$

the values given above for the variables involved, we find that on the west coast $\delta G/\delta S \sim -0.080$ m^{3}ha^{-1}a^{-1}/cm and $\delta G/\delta P \sim -0.0032$ m^{3}ha^{-1}a^{-1}/ –°Cd while in the Kainuu region $\delta G/\delta S \sim -0.035$ m^{3}ha^{-1}a^{-1}/cm and $\delta G/\delta P \sim 0.0013$ m^{3}ha^{-1}a^{-1}/ –°Cd.

Consequently, an enhancement of the mean snow depth in winter by 20 cm under an unchanged frost sum, or a reduction of the frost sum by 500 ((–°C) d^{-1}) under an unchanged snow depth causes an increase in the growth of forest stands by 0.7–1.6 m^{3}ha^{-1}a^{-1}.

For evapotranspiration in the forests we have

$$\delta Ef/ \delta S = \delta Ef/ \delta G \bullet \delta G/ \delta S \sim 30\ \delta G/ \delta S \text{ mm cm}^{-1} \tag{30}$$

$$\delta Ef/ \delta P = \delta Ef/ \delta G \bullet \delta G/ \delta P \sim 30\ \delta G/ \delta P \text{ mm } (-°\text{C}^{-1}\ \text{d}^{-1})\ . \tag{31}$$

For example, let us enhance precipitation during the winter by 40 mm with an unchanged frost sum so that the snow depth on the west coast increases by 15 cm and in the other regions by 20 cm. As a consequence, the annual evapotranspiration on the west coast increases by 36 mm and in the Kainuu region by 21 mm. The annual runoff thus only increases by 5–20 mm. For Lapland one should also note that an increased snow cover in the forests enhances the area covered by water on aapa fens and so the evapotranspiration from them.

Assume an increase of 500 $(-°C)^{-1}$ d^{-1}) in the frost sum, corresponding to a decrease of 4 °C in the winter-time mean temperature, and an unchanged snow depth. As a consequence, annual evapotranspiration decreases and runoff increases on the west coast by 48 mm and in the Kainuu region by 20 mm.

Assume that the winter's greatest soil frost depth increases by 15 cm. As a consequence, annual evapotranspiration, calculated as $\Delta E = \delta E/ \delta G \bullet \delta G/ \delta F \bullet \Delta F$, as 30 • 0.0435 • 15, decreases by 20 mm.

All these responses are actually distributed over a few preceding years, with an increasing weight forwards in time. These examples thus apply best for means over several years.

5. Conclusions

The temporal standard deviations of the annual variation of evapotranspiration and forest growth were derived as functions of climatic variables for an area covering most of Finland and including both the southern and middle boreal ecoclimatic zones. In particular, the roles of soil frost, snow depth and frost sum were taken into account, and the main regional features were brought out. The verification provided by Class–A pan observations and the results from tree-ring studies showed that the temporal variations were of the right magnitude considering that the response of the growth of forest stands to climate is distributed over several years.

The results of this study, in spite of their rather preliminary nature, should encourage climatologists, hydrologists and forest researchers to increase interdisciplinary cooperation in this field. The next step would be the explaining of observed annual evapotranspiration and forest growth records by the annual values of the effective climatic variables. Note also that in the near future, a series of over 10 years of data for the annual variation of evapotranspiration will become available on the basis of flux studies.

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