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Swedish Evaporation Research
A Review

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Swedish evaporation research has gone through a rapid development, and there is now a need to overview the results. This review presents the research in a broad sense. The brilliant research by Wallerius between 1739 and 1747 clarified the basic nature of the evaporation process. After him, research has been concentrated to two periods. The first, in the beginning of this century, was connected with the establishment of the meteorological and hydrological services in Sweden. The second period of progress started with the International Hydrological Decade (1965-1974). Recent evaporation research has been strongly focused on forest, and aimed at establishing the causal relationship between changes in land use and changes in the hydrological cycle. Applications to problems of clearfelling, large-scale introduction of water-demanding energy forests, and effects of atmospheric pollution have shown the great importance of the evaporation component of the water budget. Present evaporation research in Sweden is directed along three main lines. Firstly the development of reliable methods to measure evaporation. Secondly sophisticated turbulence measurements within and immediately above a vegetation cover. Thirdly the development of simple, yet physically realistic models of evaporation from plant covers.

Introduction

Evaporation is the most difficult term to assess of all components in the water balance. It is also the most important term for several problems in, *e.g.*, hydrology, agronomy, forestry or land resource planning. Since development in the last decades has considerably improved both quality and quantity of data and theories concerning evaporation, there is now a need to present an overview of this research.

Reviews of Swedish evaporation data, methods and research are scarce and of variable extent and quality. A. Wallén (1911) reviews the old history of Swedish hydrology from the earliest times until 1870, including the efforts devoted to evaporation. In two later papers, Wallén (1914, 1918b) presents the different methods available for determination of evaporation, with a strong emphasis on lake evaporation. Wallén's reviews correspond to important efforts in evaporation research that followed on the establishment of the first organized networks for measurement of hydrological and meteorological data in the first decades of this century. Similarly intensive efforts were not again undertaken until the advent of the International Hydrological Decade (IHD 1965-1974). One of the first evaporation related reviews after Wallén's is, consequently, by Forsman (1969), who presents a multitude of methods for measurement and calculation of evaporation, and discusses their relevance to the Swedish IHD research programme. Odin (1976), in his PhD thesis, reviews both the theory of latent heat exchange, the methods for measuring it, and the available formulæ for evaporation calculations, with special emphasis on forestry. Grip (1978) gives a similar presentation of various evaporation formulæ that could be useful for research efforts within the ten-year long Swedish Coniferous Forest Project. Within the framework of the International Hydrological Programme (IHP, started in 1975 as a continuation of IHD), Rodhe (1979) gives an overview of different methods of calculating lake evaporation with the aim of finding a simplified method of general applicability. As part of an effort within the IHP to encourage cooperation between ecologists and hydrologists, Grip (1984) reviews the interception evaporation from forests, giving both theories and data for Swedish conditions.

No review has previously presented evaporation in a broad sense. The present review covers the development of Swedish evaporation research from the earliest times. It also covers a wide variety of aspects on evaporation. This means that presentations are superficial rather than penetrating. More specialised aspects, such as human or animal perspiration or elaborate details on stomatal functioning have been left out. Regional water balance estimates of evaporation and the influence of forests on the water balance have long dominated Swedish research and so occupy a major part of this review.

Evaporation terminology is often confusing, as shown by Stanhill (1973). *Evaporation* is used in this paper for all types of water vapour transport into the atmosphere. The long and tedious word *evapotranspiration* is avoided. Transpiration signifies water vapour lost from leaves/needles after passage of the water through the plant. If evaporation emanates from, *e.g.*, soil or intercepted water this is mentioned explicitly. *Interception* signifies the process of water being caught on a vegetative cover, not evaporation of this water.

Early History

The first traces of modern hydrology are found in the beginning of the eighteenth century (Wallén 1911). Specific evaporation research starts with the investigations undertaken in Uppsala by Wallerius (1740, 1746a,b; 1747a,b). At his time very few things were obvious with regard to the nature of evaporation. The brilliant research by Wallerius clarified, *e.g.*, that water vapour originates from fluid water, that the evaporation rate is positively correlated to heat exposition and to the surface of a water body. Wallerius also gave quantitative relations between evaporation on the one hand, and insolation and wind speed on the other hand. After Wallerius very little is to be found specifically about the evaporation process until modern times. Some of the first attempts to estimate regional evaporation are discussed by Lagerheim (1842). Estimates of evaporation as a residual in water budgets are published with increasing frequency from the middle of the eighteenth century onwards. These estimates are well presented by A. Wallén (1911) and, for the period before 1823, in detail by von Ehrenheim (1824), who also presents the series of "atmidometer" measurements from the astronomical observatory in Stockholm. These measurements started in April 1786 but regular notations seem to have stopped in 1816. Regular, daily notations were published again at least from 1888, by which time the instrument had been replaced by a Wild evaporimeter, a small weighing pan commonly used around the turn of the century. This homogeneous series is still going on (personal communication, Bertil Eriksson 1982; 1988).

Regional Mapping of Evaporation

Evaporation as a Residual in the Water Balance Equation

The most reliable absolute data are obtained from calculations of evaporation, E, as a residual ($= P - A$, where P is precipitation and A runoff) on a large catchment scale. This calculation assumes that storage terms are negligible and is normally valid only for periods of several months or years.

In modern times, the basic investigations on the geographical distribution of evaporation were carried out during the 1910s and 1920s by A. Wallén. He reports a ten-year investigation of the hydrology of the Lagan catchment in southern Sweden (Wallén 1918a), for which the annual evaporation was 368 mm. In a subsequent series of papers, Wallén (1923, 1924b, 1927) gives more and more details from different catchments. In 1934 he sums up investigations from a total of 24 large catchments in southern and central Sweden. He states that the average evaporation for these parts of Sweden is 360 mm per year with extremes of 301 mm per year and 421 mm per year. His findings are summarised in the following regression equation, although he cautions the reader that it is based on a quite scattered material

$$E = 0.82\,S + 22.8\,T + 63 \quad (1)$$

E is annual evaporation in mm, S the lake percentage and T the average air temperature in °C during the growth season, defined as lasting from May to October.

A fairly dense network of runoff and precipitation gauging stations made evaporation calculations fairly trustworthy in southern and central Sweden after the 1920s, but the lack of data from the high mountain areas in northern Sweden was a major problem. A. Hamberg (1901) reports measurements made in the Sarek mountains with an evaporimeter designed by H. E. Hamberg (1885), but his results can only be regarded as non-conclusive. Wallén (1923, 1924a) suggests, on the basis of insufficient data, that high mountain evaporation amounts to 250-300 mm per year. Melin (1943) partly solved the problem of high mountain evaporation by his water balance studies in Lake Malmagen catchment. He considerably lowered earlier estimates with his investigation and shows the high mountain evaporation to be 100-150 mm per year. Melin's results are used by C. C. Wallén (1951) in his publication of a new Swedish precipitation map. Bergsten (1954) also combines the earlier results of A. Wallén and Melin in the publication of one of the first maps of Swedish evaporation (Fig. 1a). Generally speaking, one may say that these early investigations point towards a regional evaporation in southern and central Sweden as being more or less constant with a value of 360 mm per year.

Bergsten (1950) made a thorough investigation of six well studied catchments, evenly distributed over Sweden. He shows a high correlation between annual evaporation and summer temperature. Since the above-mentioned value of 360 mm was deduced from a period of rather cool summers, Bergsten suggests that this figure be raised by 10%.

Tamm (1954) reappraised the available water balance data to find a relation between the annual land surface evaporation, E, and the mean annual air temperature, T. After disregarding the high mountain areas and the great lakes, Tamm presents a good correlation with

$$E = 221.5 + 29.0\,T \quad (2)$$

where E is in mm per year and T in °C. With this relationship, Tamm's new map of Swedish evaporation (Fig. 1a) better reflects the topographical and climatic influences on the actual evaporation. Tamm (1959b) also uses Eq. (2) in a special study of the superhumid climate in southwestern Sweden describing the large local variations of evaporation in this area.

Fifteen small, "representative" catchments in the Nordic countries (of which seven in Sweden) were intensively studied and instrumented in the beginning of the 1970s as a part of the IHD. All the major components of the water balance were measured with a resolution in time and space which was higher than the previously cited studies. As an example of this research, one can mention the studies in

Värpinge, southern Sweden, where the water balance was measured in a small (1.95 km) agricultural catchment close to Lund, with the original aim to study effects of urbanisation. Background measurements between 1971 and 1980 could never be used as intended, since the Värpinge area never was urbanised. During the whole period, the catchment was used to cultivate mainly barley, wheat and sugar beats. As an average for the ten years, evaporation, given as a water balance residual, amounted to 441 mm per year, with extremes of 542 mm in 1971 and 403 mm in 1972. Average precipitation amounted to 589 mm per year with extremes of 725 mm in 1980 and 459 mm in 1975. It is interesting that evaporation during winter months was seldom much below 10 mm per month (Lindh *et al.* 1983; Hogland 1986). The aim to compare an urban and a rural area is met by the study of Hogland (1986), who compares the water balance of the city of Lund with results from the Värpinge investigations. The major difference between town and countryside is a larger runoff from the town, which reduces evaporation as a percentage of precipitation from 65 % in Värpinge to 45 % in Lund.

The accuracy of the residual evaporation estimate is no better than the accuracy of the measured components in the water balance equation. The most critical assumption in this context is that precipitation records have no systematic errors. Whether this is a realistic assumption or not, was more and more questioned during the 1960s and the 1970s, as a result of studies carried out during the IHD in the "representative" basins. Evaporation was measured intensively in these basins both by direct and by indirect methods, such as micrometeorological methods, various methods based on evaporation pans, soil water budgets, etc. These measurements were complemented with calculations based on different empirical formulæ. The IHD studies indicated an annual evaporation more than 100 mm larger than was previously assumed, *i.e.,* more than 20 % larger in southern and central Sweden (see, *e.g.,* Waldenström 1977).

The best available evaporation estimates have, thus, been achieved by a careful study of old precipitation records. Eriksson (1980b) analyses such records from the last two centuries and shows long-term variations of such an amplitude and cycle that the concept of the climatic "normal period" 1931-1960 should be regarded with scepticism. Eriksson (1980a) reappraises older estimates of regional precipitation. He accounts for losses from the precipitation gauges caused by wind, evaporation and adhesion. From precipitation data, separately corrected for each of his 260 stations, he concludes that older precipitation estimates must be increased by 28 % as an average for all of Sweden. Out of these, 18 % result from gauge loss corrections, whereas 10 % result from the uneven distribution of gauges at different elevations. The map (Fig. 1a) presented by Eriksson (1980a) should, thus, be regarded as the present state-of-the-art.

Alternative Methods to Determine Regional Evaporation

Although regional evaporation has been given almost exclusively as a water ba-

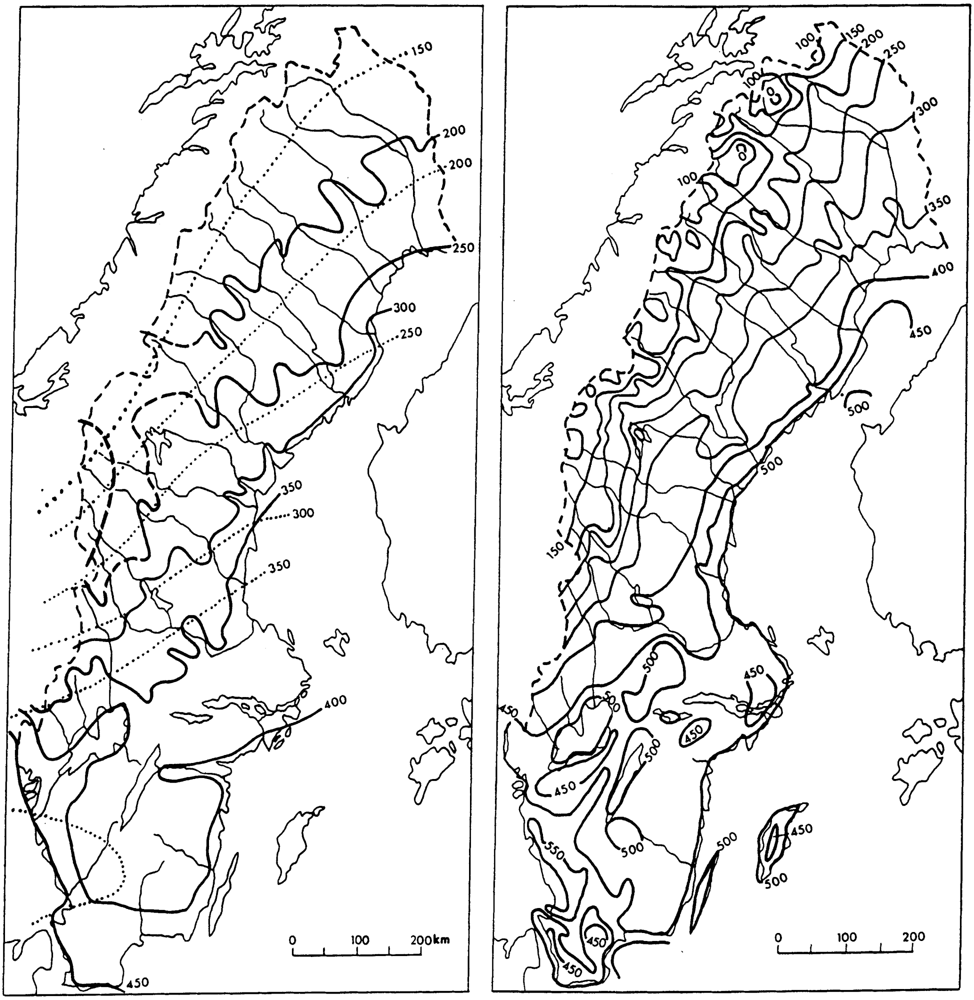

Fig. 1a. Actual evaporation (mm/year) over Sweden calculated as a residual water balance term. The left map, redrawn from Tamm (1959), shows results by Bergsten (1954) as dotted isolines and by Tamm (1954) as solid isolines. Both authors based their calculations on data from 1921-1950 and did not correct for precipitation errors. The right map, redrawn from Eriksson (1980a), is based on data from the period 1931-1960 and on corrected precipitation records.

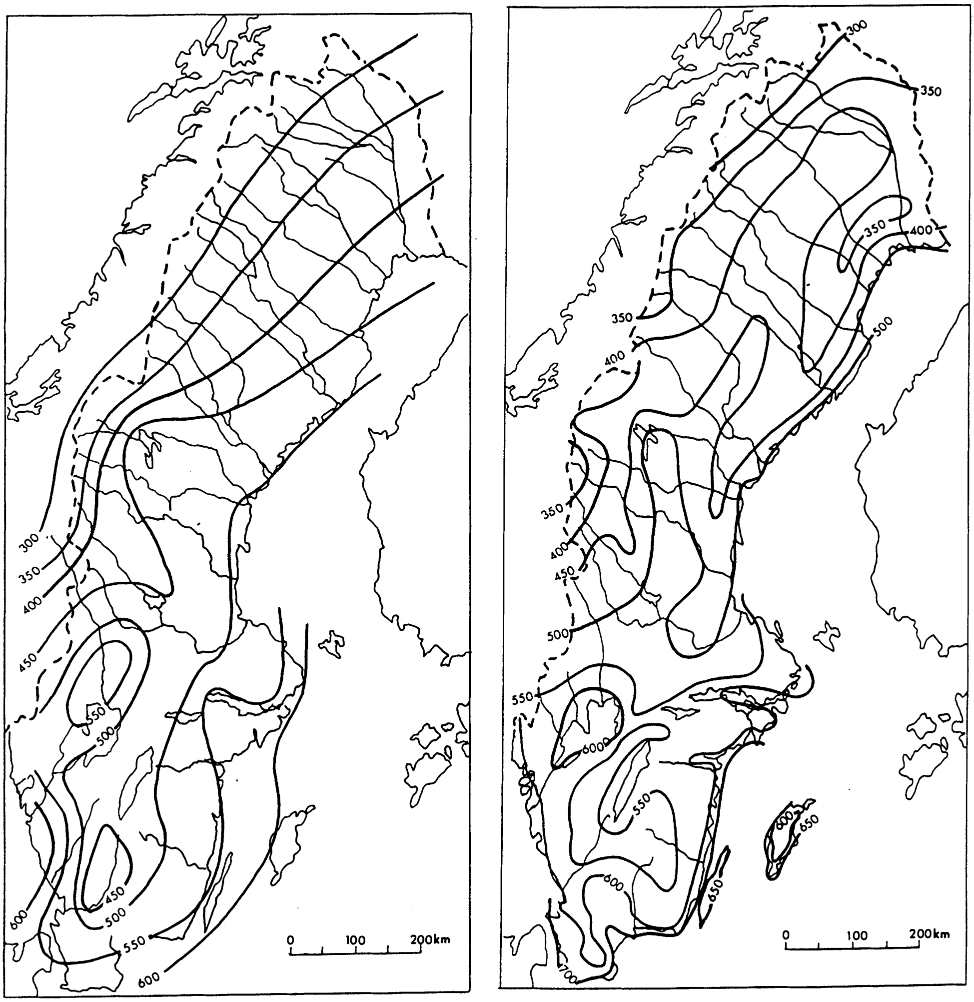

Fig. 1b. Potential evaporation (mm/year) over Sweden calculated with the Penman (1948) formula. The left map, redrawn from Wallén (1966), is based on data from the period 1931-1960, whereas the right map, redrawn from Eriksson (1981), is based on data from the period 1961-1978.

lance residual, some other methods have been occasionally used. Wallén (1929), for example, attempts to calculate evaporation in central Sweden with an energy balance method. He reports evaporation values of the same order as those from water balance calculations, but he also describes several problems with the method.

Nyberg (1965) reports evaporation measured with the aerological method in southern Sweden during 1957. Readings from radiosondes, taken twice a day at eight different places, were used to calculate the net flux of water vapour over the delimitations laid out over an area encompassing southern Sweden. The aerological method supplies information on the difference between P and E, and even if Nyberg uses precipitation data from more than 400 stations to reduce the error caused by insufficient areal representativity, he presents his result as

$$E = 380 + p \pm 50 (\mathrm{mm/year}) \tag{3}$$

Here p is the systematical error in precipitation data. With this correction equal to zero, Nyberg's (1965) results agree well with earlier water budget studies.

No results from investigations in Sweden using the aerological method have been published after Nyberg (1965), but it is suitable to mention the studies by Alestalo and Savijärvi (1977). They attempted to apply the method to northwestern Europe, but conclude that reliable results can only be obtained over fairly large, homogeneous areas, such as the Bothnian Sea.

Practically Useful Estimates of Evaporation

Although the residual evaporation estimates give reasonably accurate values for long periods and large areas, they are not well fitted to meet the demands of, *e.g.*, foresters and farmers. In order to calculate the water available to vegetation, and to relate plant growth to climate, it is necessary to know the evaporation for periods much shorter than years and for areas of smaller extent. Since such measurements were hardly available in the beginning of this century and are only scarcely available today, two practically useful concepts have been developed.

The first concept is that of a humidity index, which does not necessarily have to explicitly incorporate evaporation. Hesselman (1932) presents several indices proposed in the beginning of the 1930s, and suggests de Martonne's (1926) index to be best suited to Swedish conditions. This index is defined as the quotient $P/(T + 10)$, where P is annual precipitation in mm and T is the mean annual air temperature in °C. Tamm (1954, 1959a) argues against this index and instead suggests that humidity $H = P\text{-}E$, where E equals the mean annual evaporation, is the best choice. Tamm suggests that his previously mentioned formula, based on mean temperature only, be used to calculate the evaporation E needed for this index. Ångström (1958) compares Tamm's formula (Eq. (2)) and de Martonne's index and concludes that, for Swedish conditions, the latter index equals the quotient P/E which is presented by A. Wallén (1927) as a suitable humidity index.

The second concept is that of potential evaporation, a term coined by Thornthwaite in 1948. This concept is based on the idea of an evaporation determined exclusively by atmospheric demand. Although this idea has been shown to be unrealistic and erroneous in numerous applications, the potential evaporation concept has gained a widespread acceptance as a practically useful tool, not only in forestry and agriculture, but also in hydrology and civil engineering. The combination equation of Penman (1948), in one form or another, has turned out to become a *de facto* standard for the calculation of potential evaporation both in Sweden and internationally. One of the problems with Penman's equation is that it requires information on solar radiation which is not commonly measured. C. C. Wallén (1966) uses eight stations to establish a relationship between global solar radiation and sunshine hours and then uses sunshine hours from 35 stations to map the global radiation for the period 1931-1960. With these data, he calculates the Penman (1948) evaporation and presents the first map of this in Sweden (Fig. 1b). Requests, especially from hydrologists, to improve Wallén's results by inclusion of more stations and by calculations of monthly values of the potential evaporation, caused Eriksson (1981) to publish a new map (Fig. 1b) of this entity. It is noteworthy, though, that Eriksson finds the entire concept to be of restricted value and recommends his readers to use his values with the utmost care.

Even if the potential evaporation concept has largely replaced the humidity index in most applications, it can not give any direct estimation of the amount of water available for plant growth. To meet this demand for more applicable climate data, Eriksson (1986) presents new maps of Tamm's humidity index for the whole growth season as well as for parts of it. As input to his calculations he uses precipitation data corrected for systematic errors in combination with both potential and actual evaporation. For the potential evaporation he uses the Penman (1948) formula and for the actual evaporation he applies gross multiplicative factors, variable with latitude, to the potential evaporation. Tamm's humidity index is also used in a study to evaluate the impact on water resources of water demanding energy forests which will probably be introduced on a large scale in Sweden in the beginning of the 1990s (Lindroth and Halldin 1988 and Halldin and Lindroth, in press).

Forest Evaporation

Early Investigations

Since a major part of Sweden is covered with forests, it is natural that forest evaporation early attracted a large interest. The first major work in this field was by Hamberg (1885, 1889, 1896a,b), who carried out a thorough investigation that eventually formed the basis of today's synoptic network in Sweden. The reason to study forests in connection with water budgets was twofold. Firstly that water availability was seen as a major determinant for forest productivity. This aspect primarily attracted forestry researchers, and Tamm (1955) has reviewed this work. The second aspect, namely the influence of forests on the water quality and runoff, has attracted equal interest from hydrologists and forestry researchers. Stålfelt (1944) was the first to make a direct assessment of forest evaporation. In a small spruce stand in southern Sweden he investigated transpiration loss by the "momentan" method, *i.e.,* by cutting shoots and immediately weighing their decreasing mass. In combination with other measurements he was able to give a fairly complete description of the water balance in his young spruce stand during a period of four years. For this period, the well supplied trees transpired 378 mm during the growth season. Stålfelt's conclusion is that the yearly precipitation was not sufficient to supply the spruce with the necessary water and that the forest, in this case, acted as a drain. Stålfelt's fidings seriously challenged the established hydrological view. In reaction to this, Tryselius (1946) states that spruce evaporation can never exceed 400 mm. Stålfelt (1947) maintains his view and the query was never settled. As an offshoot of this debate, Kihlberg (1958) presents results from the first Swedish investigations carried out on the effect of clear-cutting. On two small, almost identical, forested catchments at the Himmelsberget mountain, runoff and precipitation were measured from 1925-26 to 1930-31. At this time the western catchment was clearfelled, after which measurements were continued until 1936-37. Evaporation from the western catchment went down from 369 mm to 305 mm per year, in spite of the fact that evaporation from the untouched, eastern watershed went up from 318 mm to 350 mm per year. Kihlberg (1958) concludes that the presence of a forest cover certainly gives rise to an increased evaporation, but hardly to the degree that the forest can start to act as a drain.

What Causes Forest Evaporation?

Modern forest evaporation research in Sweden starts with the IHD period and has been concentrated on finding results of general applicability in order to address questions like, *e.g.,* how changes in land use may affect evaporation. Investigations have been carried out primarily at two sites. Firstly in a mixed forest in Velen, southern Sweden, between 1971 and 1974. Secondly in a pine forest in Jädraås, central Sweden, from 1976 and still continued (the time of writing is 1988).

It was an aim with the Velen measurements to compare different measurement methods, and also Jädraås data have served to evaluate estimation methods. All these estimations have pointed at the severe problems related to evaporation measurements above forest canopies. The central aim with all these late measurements was to find the mechanisms that regulate forest evaporation. As we have seen earlier, widely different results had been achieved from different forests at different times. In agreement with the findings of Rutter (1963), both Velen and Jädraås data strongly support the view that evaporation rates from an intercepted forest canopy are much higher than transpiration rates under the same circumstances (Bringfelt *et al.* 1977; Lindroth 1984b, 1985b). Data from Velen and Jädraås further support the view that even under conditions of ample water supply, forest transpiration is regulated by stomatal control rather than by climatic demand. The study of forest evaporation, therefore, to a large degree becomes the study of the surface resistance of the forest canopy (Halldin *et al* 1980; Bringfelt 1980).

Problems related to measurement of forest evaporation

Direct measurements of evaporation from forests are specific in at least four respects. Firstly since large practical problems arise when one wants to perform measurements at 20 m or more above the ground. Secondly in that the gradients of temperature and humidity are commonly extremely small above a forest canopy, and thirdly in that a sufficient fetch requires a horizontally homogeneous forest of an extent that hardly exists in Sweden. A final aspect that makes forests more difficult to treat than lower vegetation, is their heat storage properties. Bringfelt and Orrskog (1976) measured heat storage rates in the Velen forest and present 20 W/m^2 as a maximal value into each of the soil, the biomass and the canopy air compartments. Such heat flux rates may be of the same order as the latent heat flux constituting evaporation in certain circumstances. Lindroth (1985b) reports average daytime heat storage fluxes in the Jädraås forest to be 23 W/m^2 during May and September.

Aside from water balance measurements, the major micrometeorological techniques (presented by Bringfelt 1975) were tested in Velen in a 54 m high mast. One of the earliest discoveries was that the aerodynamical profile method was very difficult to apply. In this method, three parameters (u^*, d, z_0) must be determined simultaneously, and this determination turned out to be an ill-defined problem. The method was, thus, abandoned in favour of the eddy correlation and the energy balance Bowen ratio techniques (Bringfelt *et al.* 1977). As a consequence of the Velen results, among others, the micrometerological measurements in the sparse pine forest in Jädraås made use of only the energy balance Bowen ratio method (Perttu *et al.* 1977, 1980; Lindroth and Norén 1979).

Bringfelt (1980) reports that Bowen ratio estimates of evaporation in Velen are consistently larger than the corresponding eddy correlation estimates. Abandoning the similarity principle and putting the quotient of eddy diffusivities for heat and vapour (K_h/K_e) equal to 1.3 would cure this deficiency but would, on the other hand, cause a considerable discrepancy in the estimations of sensible heat exchange. Comparison during seven months of 1973 and 1974 of evaporation estimated by the water balance ($E_{\mathrm{w.b.}}$) and the Bowen ratio (E_{Bowen}) methods also shows a consistent difference, with the Bowen ratio estimate always being the higher. For one particular month the quotient $E_{\mathrm{Bowen}}/E_{\mathrm{w.b.}}$ was close to 200 %, but the average was closer to 130 %.

Comparison of evaporation estimates by the water balance and Bowen ratio methods at Jädraås is reported by Grip *et al.* (1979) for the major part of the 1977 growth season. The result in this case is even more baffling than the one from

Velen, with energy and water balance estimates of evaporation being 350 mm and 150 mm, respectively. This discrepancy is so significant that theoretical explanations must be sought. Grip *et al.* (1979) propose that one such explanation can be that the similarity principle is not valid over rough forest surfaces. Lindroth (1984a) analyses the gradients above the Jädraås forest and shows that there exists a layer immediately above the crowns, the roughness sublayer, in which temperature and humidity profiles are disturbed. Lindroth recalculates the Bowen ratios from Jädraås without data from this sublayer. This way he decreases the energy balance estimate from 350 mm to 300 mm. The discrepancy has, thus, not yet been resolved.

Surface resistance of a forest cover

The surface resistance, r_s, of a vegetation cover is defined by the Penman (1953) equation (commonly and erroneously attributed to Monteith (1965) see footnote, page 8 in Monteith 1981)

$$LE = \frac{\Delta R_n + \rho c_p \delta e / r_a}{\Delta + \gamma(1 + r_s / r_a)} \tag{4}$$

LE is the evaporation in mass units, Δ the slope of the temperature function of the saturated vapour pressure, γ the psycrometric "constant", ϱc_p the heat capacity of the air, δe the vapour pressure deficit, and r_a the aerodynamic resistance. It is implicit in the papers by Monteith (1964, 1965) that the resistance of all stomata in parallell equals the surface resistance, but this assumption has been severely critized in the litterature (see, *e.g.,* the discussion on pages 111-112 in the end of Monteith's (1964) paper). The determination of a surface resistance can be split up in three parts. Firstly the determination of the stomatal resistance of a leaf or a needle, secondly the determination of the area of all leaves/needles, and thirdly the determination of the relation between the surface resistance and the two previous entities.

One important result of both the Velen and the Jädraås investigations is the presentation of data that strongly support a functional form for the biological control of transpiration. The Lohammar equation, which represents a semi-empirical deduction of the functional relationship between climate and stomatal conductance, k_s, is presented in two papers of Lohammar *et al.* (1979, 1980)

$$k_s = \frac{R_g}{R_g + R_o} \frac{k_{max}}{1 + bVCD} \tag{5}$$

R_g is global shortwave radiation, *VCD* vapour concentration deficit, whereas R_o, k_{max} and b are constants. On the presumption that the surface resistance, r_s, equals the resistance of all stomata in parallell, one can write

$$r_s = (k_s LAI)^{-1} \tag{6}$$

where *LAI* is the leaf/needle area index, *i.e.,* the area of the total leaf/needle surface per unit ground area.

Eqs. (5) and (6) have been used successfully by Halldin *et al.* (1979) in a model to predict evaporation from the Jädraås forest. Later on, Bringfelt (1982a) used the Bowen ratio measurements from Velen to test the inverse of Eq. (5) as a direct expression of the surface resistance. After comparing various functional expressions he found the Lohammar equation to give the best agreement with measurements. Lindroth (1985a) used data from Jädraås to calculate a specific canopy conductance, scaled by the *LAI* of the canopy. He also found the Lohammar equation to give a very good fit.

Lindroth and Halldin (1986) analyse climatic profiles inside the Jädraås forest canopy and show that surface resistance can not be scaled directly by the *LAI* to give an inverse value of the stomatal conductance. The same conclusion is reached by Bringfelt (1986) and Bringfelt and Lindroth (1987) who compare parameter values for the Lohammar equation (directly applied as a surface resitance equation) from both Velen and Jädraås and show that scaling with *LAI* does not make the values coincide.

One prerequisite for relating stomatal conductances and the surface resistance of a stand is that the total size and the spatial distribution of the leaf/needle area be known. Halldin (1984) presents a simple method to calculate the vertical distribution of biomass elements in the Jädraås forest, based on simple forestry measures only. With this method, Lindroth and Halldin (1986) analyse the relation between the surface resistance and the vertically distributed stomatal conductances. They show that the calculation of the average stomatal conductance of the canopy is crucial to the success of Eq. (6). A correct averaging requires values from all heights in the canopy, preferably weighed by the leaf/needle area density at each height. Lindroth and Halldin (1986) also show that Eq. (6) is a reasonably valid approximation if applied only to the part of the evaporation that emanates from the canopy, *i.e.,* that evaporation from the forest floor must be treated separately.

Another prerequisite for the operational application of Eqs. (5) and (6), is that the variation in stomatal control within or between seasons must be known or negligible. Lindroth (1985a,b) presents the seasonal variation of the energy balance components from Jädraås for 1977 and 1978. He shows that the Bowen ratio at noon decreases systematically from spring to autumn, *i.e.,* that more of the available energy is spent on evaporation towards the end of the growth season and less on heating the air. He is also able to explain the seasonal variation in the surface resistance by three factors. Firstly that the seasonal variation in climate explains 50 %, secondly that the variation in *LAI* explains 30 %, and finally that 20 % is explained by variations in the stomatal efficiency. Most of the change in stomatal efficiency takes place in spring and early summer and can possibly be explained as a result of remaining physiological effects from the hibernation. Lindroth (1985c) shows that such variations in needle area within and between seasons give rise to variations of much smaller size in the total evaporation from the stand. The influence of an increase or a decrease in needle area of 25 % is reflected in a change of total evaporation of less than 10 %, whereas the interception evaporation changes by 20 %.

Interception evaporation

One climatic effect of a forest canopy relates to its capacity to intercept precipitation. As remarked earlier, there is a general concensus today that interception evaporation must be treated as a separate part of forest evaporation, and this makes interception studies important for evaporation research. The first interception studies in Sweden were carried out by Hamberg (1896a,b). With results from rain gauges in the open and under the canopy, he concludes that between one quarter and one third of the incoming precipitation was lost as interception evaporation. For young spruce he mentions a saturation capacity of 4 mm. He also concludes that the lighter the rain, the larger the fraction of it that is evaporated from the canopy. As an odd detail one may mention that he found spruce to intercept snow less than rain but for pine the opposite was true. After Hamberg, little progress was made before Stålfelt (1944, 1963) carried out his thorough study on the water balance of spruce. Stålfelt constructed a special trough for measurements under the tree crowns, whereas he used ordinary cylindrical vessels in the open spaces between trees. He filled some of the troughs and vessels with undisturbed samples of the ground and litter layers, thereby attempting to measure ground layer interception. With this setup he realized the large difference between amounts of rain collected in glades and on open fields. This he attributed to the "rain-shadow" of the tree. With his investigation he was able to quantify interception in crowns and litter layer both for the growth season and for the remaining part of the year. On average 30 % of the precipitation reached the root zone from May to August, whereas 41 % reached it during the dormant season.

In connection with the IHD programme in Velen, Bringfelt and Hårsmar (1974) measured interception during the summer and autumn of 1973. They analyse the data according to the theory by Rutter *et al.* (1971/72). For the mixed pine-spruce-birch forest in Velen the free throughfall coefficient was estimated to around 0.5 both from interception measurements and from fish-eye photographs. The storage capacity was judged to be 2 mm. With the Velen precipitation climate, this means that 26 % of the precipitation was lost as interception. Results of interception measurements in the Jädraås pine forest are reported by Perttu *et al.* (1980). Eriksson and Grip (1979) use these data to compare three different interception models. They report free throughfall coefficients of 0.65 for a young, a dense and a mixed spruce-pine stand as well as for a mature pine forest. Storage capacity for the old forest was only 0.7 mm. Stemflow was negligible and a total of 20 % of precipitation was lost as interception evaporation during summer. Perttu *et al.* (1980) also make an attempt to deduce the momentary drip function based on a formulation modified from the one given by Rutter *et al.* (1971/72). Their hope was to find a variation of the function parameters with wind speed, but the quality of their data did not allow this.

Grip (1981) reports interception measurements in energy forestry *Salix* stands in Studsvik, central Sweden between 1978 and 1980. This kind of forest has properties completely different from those of coniferous forests. A surprisingly high percentage (33 %) of the precipitation got channeled as stemflow, whereas throughfall accounted for between 31 % and 56 %. Interception evaporation, thus, varied between 36 % and 11 % of precipitation. Andersson (1987) presents data for a willow stand in Uppsala, central Sweden. Measurements were carried out of stemflow, throughfall and interception evaporation and Anderson (1987) relates these to surface properties. Stemflow varied between 3 % and 19 % of precipitation, throughfall varied between 74 % and 45 %, and interception evaporation between 2 % and 36 %. It is curious that the interception evaporation percentage increased during the growth season for Andersson (1987), whereas it decreased for Grip (1981). Andersson's (1987) data show that the canopy routes the intercepted water to stemflow more efficiently the larger the leaf area, *i.e.,* that the willow stand partly acts as a funnel.

One of the major elements in the generally accepted interception theory of Rutter *et al.* (1971/72) considers the relation between canopy storage, C, and evaporation rate, E

$$E \equiv C/S\, E_0 \tag{7}$$

where S is storage capacity and E_0 evaporation from a completely saturated canopy. Rutter *et al.* base their relation on rather vague data, but Larsson (1981) is able to experimentally confirm the validity of Eq. (7). Larsson weighed small stem segments of *Salix* placed with their lower ends in sealed water beakers. In this way he separated interception evaporation from transpiration and demonstrates the way they influence each other. Not only does he confirm the relation of Rutter *et al.* (1971/72), but also a relation for the combined transpiration and interception evaporation

$$E = (1-C/S)T_0 + C/S\, E_0 \tag{8}$$

where T_0 is potential transpiration.

Evaporation from the forest floor

None of the Velen and Jädraås studies paid much attention to the role of the ground layer. This, on the other hand, is an important topic in Odin's (1976) investigations on how forest edges affect the local climate. He studied the wind field and the evaporation in forests and clear-felled areas at four sites in the far north of Sweden in the 1960s. At one clearfelling, Odin used the aerodynamical profile method to determine the energy balance components. Curiously enough he found the evaporation to decrease slightly with wind speed. In a later study, Johnsson and Odin (1978) present an accurate dynamic weighing lysimeter. With this, values of evaporation, transpiration, condensation or fog deposition can be recorded automatically every 1-2 minutes. The lysimeter was tested for two days in October 1976 on lichen and heather vegetation at a clearfelling, and its readings are compared with Bowen ratio estimates. In this case the Bowen ratio estimate is more than twice the lysimeter value, the discrepancy being attributed to advective effects. Lindroth (1984b) used the same lysimeter in Jädraås to find out that evaporation from the forest floor amounted to 12 % of the stand evaporation for this rather open pine stand.

Models of Forest Evaporation

Since no single formula is sufficient to describe or predict the evaporation from a forest, there have been several efforts to create models of various complexity to simulate the process. Halldin *et al.* (1979) and Halldin *et al.* (1980) discuss several models for the whole or parts of the forest evaporation process. The synthesis of these different modelling efforts is a simple, yet physically realistic model, given the name KAUSHA (Swahili for evaporation), and originally presented by Halldin and Grip (1979). This model separately treats the transpiration and interception evaporation and accounts for soil water influences by a simple budget. The basis for the transpiration sub-model is the Lohammar equation. One special problem addressed by the KAUSHA model, is that of modelling instantaneous processes like transpiration with the use of climatic input data available only once a day. In the KAUSHA model this problem is solved by dividing the 24 hours into separate day- and night-time parts, during each of which the climate is assumed constant (Halldin 1988). The same approach to transpiration modelling is used in a model presented by Bringfelt (1982b), which differs from the KAUSHA model in its more refined treatment of the interception evaporation. The problem of inconsistent time scales for modelled processes and input data is given another solution in a quite detailed model of water movement in the soil-plant-atmosphere continuum, devised by Kowalik and Eckersten (1984) for use in willow forests. Eckersten (1986) presents a method to deduce 10-minute data from synoptic weather information. The synthetic 10-minute data are then used as input to the model. This method gives output that correlates well with output deduced from measured data, except under cloudy conditions.

Because of its high demand for structural input data, the model of Kowalik and Eckersten (1984) has not been widely applied. The KAUSHA model, on the other hand, has been applied to several forests and problems. Besides its original application to the Jädraås forest (Halldin and Grip 1979), it has been used, *e.g.,* to predict the water balance of oak and beech forests in France (Halldin *et al.* 1984; Saugier *et al.* 1985) and a wet-land willow forest in Jädraås (Grip *et al.* 1984). It has also been used to evaluate willow stand lysimeter measurements in Studsvik, central Sweden, in order to arrive at a set of parameter values in the Lohammar equation, generally applicable to Swedish *Salix* forests (Grip *et al.* 1988). Bringfelt's (1982b) model has been applied to data from several synoptic stations (Bringfelt 1982b, 1985a, 1986). In the last of these papers the model is improved by the addition of empirical multiplicative corrections to transpiration in spring and in autumn. The model has also been tested on data from both Jädraås and Velen, in order to find out if parameter values in the Lohammar equation would scale with the different needle areas of the stands (Bringfelt 1986; Bringfelt and Lindroth 1987).

Does Evaporation Depend on Land Use?

One of the main reasons for modern forest evaporation research has been to establish the causal relationships between changes in land use and changes in the hydrological cycle. It is argued by Roberts (1983) that forest transpiration is a conservative process such that annual transpiration is fairly independent of species, stand density and site all over Europe. Lindroth (1985b) lends credit to Roberts' hypothesis by comparing data from Velen and Jädraås with those from Thetford in England and Ebersberger in Germany. The question remains, however, whether this conservative property is not simply an effect of either a long-term adaptation of species to a given climate or of a careful species selection by foresters responsible for reforestation of a given site. If this is the case, one might still find an appreciable influence on the water resources by deforestation or reforestation.

One case where such an influence might be considerable is in connection with the large scale introduction of water-demanding energy forestry in Sweden in the 1990s. Grip (1981) investigated the water balance of a fast growing *Salix* stand. His aim was to find out whether the soil water would be a limiting factor for energy forestry. With data from three small lysimeters, he shows the evaporation over the growth seasons of 1979 and 1980 to be 85 % of a total of 350 mm precipitation per season. Grip's (1981) data have subsequently been used by Grip *et al.* (1988) who found that, in the end of the growth season, evaporation from intensively cultivated willow stands was almost 2 mm larger per day than the so-called "potential" Penman (1948) evaporation. Lindroth and Halldin (1988) evaluate the influence that this might have on water resources in southern Sweden, and conclude that local effects may be considerable if plantations are not carefully planned.

A more traditional problem is to assess the effects of clearcutting large areas. After the initial studies by Kihlberg (1958), this problem has been addressed, *e.g.*, by Grip (1982), Rosén (1984), Grip (1987), Lindroth and Grip (1987), and Jansson (1987). Lindroth and Grip (1987) report Swedish values between 90 mm per year and 406 mm per year for the increase of runoff after clearcutting. Grip (1987) and Lindroth and Grip (1987) analyse the reasons behind this increase and show how it depends on a decrease in evaporation caused by changes in radiation climate and resistances in the air to the vapour flux.

Lindroth (1987) attacks the problem of forest damages in relation to atmospheric pollution. He uses the KAUSHA model (see above) to investigate effects on forest evaporation caused by increased needle surface of the forests. Based on the hypothesis that a forest stand must be adapted to the long term water balance of a site in such a way that it can survive an extreme drought, Lindroth (1987) shows that forest biomass growth caused by atmospheric nitrogen fertilization has considerably increased the chances for water stress. With simulations, based on 126 years of precipitation data from Falun, central Sweden, he shows that a stress that previously would occur every 50 years would occur every third year if the needle area index increased from 1.5 to 2.25 for his representative pine stand.

Bog Evaporation

The study of bogs is considered as a silvicultural discipline in Sweden, but bog evaporation has attracted very little attention. Some studies were initiated during the IHD period, but only Johansson (1974) has published any results. In the Komosse bog area in the southern Swedish highlands, he measured evaporation with two lysimeters (GGI 1000, filled with undisturbed soil cores) and several varieties of pans. Evaporation amounted to 400 mm between April and October 1972. The GGI 3000 pan gave 460 mm and the class A pan 620 mm for the same period.

Evaporation from Agricultural Land

Micrometeorological Studies

Högström and Larsson (1968) determined the water balance from 1958 to 1963 in Ugerup, a small natural catchment in a flat, sandy agricultural area in southern Sweden. Runoff and precipitation, soil and ground water contents were measured by conventional means, whereas evaporation was measured with a special technique. The vertical water vapour flux was first measured with an eddy correlation setup, devised by Högström (1967a). Fortyfive readings of this type then formed the basis for a relation between the atmospheric stability and the eddy diffusivity for water vapour (Högström 1967b). This relation was subsequently used to evaluate evaporation on a running hourly basis from continuous measurements at one height of wind speed and gradients of temperature and vapour pressure (Högström 1968). Results from this study show that the annual evaporation was 430 mm, whereas precipitation amounted to 575 mm. Since evaporation was measured directly in this case, the water balance came up with a residual of 46 mm. This is interpreted by Högström and Larsson (1968) as a systematic underestimation of the precipitation. It is also interesting to note that winter evaporation at this site was considerable (5-30 mm per month) and much higher than estimates from both the

Penman (1948) and Thornthwaite (1948) formulae, which commonly predicted net condensation during the period. This is in agreement with the results from Värpinge in the same part of Sweden, discussed in the paragraph on regional evaporation.

Högström (1974) studied the turbulent exchange of heat, water vapour and momentum over a "nonideal" agricultural site in Marsta, central Sweden. Measurements were carried out of profiles of wind speed, temperature and humidity, and of fluxes of momentum, heat and humidity by turbulence instrumentation (Smedman-Högström and Högström 1973). In unstable conditions, theoretical and measured profiles of wind and temperature agree well, assuming von Kármán's constant to equal 0.35. With strong instabilities, dimensionless profiles of vapour pressure, ϕ_e, and temperature, ϕ_h, agree well, whereas the vapour pressure closely follows the wind profile, ϕ_m, for neutral conditions, for which $\phi_m = 1.35\ \phi_h$. Högström (1974) devised a combined method to predict evaporation for all stabilities. In the neutral and unstable cases he calculates the Bowen ratio assuming that $K_h/K_e = 1.35$, *i.e.*, non-similarity of the turbulent exchange coefficients for heat and water vapour. Evaporation then follows from the energy balance. For the stable case, the energy balance is also used, but sensible heat is determined directly from the flux-gradient relationship and evaporation is given as a residual. With this method, Högström (1974) determines evaporation with an accuracy of $\pm 20\%$, based on continuous measurements on two heights only.

The Marsta data have been used to analyse the surface resistance from an agricultural field. With data from four clear summer days in 1969 and 1970, De Heer-Amissah *et al.* (1981) show that surface resistance continuously increased during the day. For two rainy summer days, the resistance fluctuated around a constant value. De Heer-Amissah *et al.* (1981) also relate the surface resistance, r_s, to the Bowen ratio, β, and to the climatological resistance, r_I (see Stewart and Thom 1973). They found that, for $r_I/r_s < 0.7$

$$r_I/r_s = m\alpha \tag{9a}$$

$$r_s = \exp(a-b\alpha) \tag{9b}$$

where

$$\alpha = 1/(1+\beta) \tag{9c}$$

Considering that r_s and β are commonly difficult to measure, whereas r_I is fairly easy to obtain from measurements, the combined equation

$$r_I = m\alpha \cdot \exp(a-b\alpha) \tag{10}$$

can be used to deduce both r_s and β. The regression constants *m, a* and *b* are given by De Heer Amissah *et al.* (1981) both for the Marsta grass-covered site and for the Thetford forest (see Stewart and Thom 1973).

Use of Evaporimenters in Agricultural Research

Swedish hydrologists and meteorologists have regarded all kinds of evaporimeters with a large dose of scepticism. Researchers in the agricultural sector have been less sceptical and found evaporimeters to fill a need, *e.g.*, as indicators of humidity. During the IHD several attempts were made to use, *e.g.*, the American class A pan and the Russian GGI 3000 pan. After the IHD period these attempts came to an end, and only Andersson's (1969) evaporimeter is used on a routine basis at present. Andersson constructed this nationally well known instrument in 1958, although he did not publish his construction before 1969. His evaporimeter is aimed at combining the advantages of small, quickly responding evaporimeters (like the Piche atmometer) and larger pans, more suited for long-term readings. The long-term usage requires the meter to contain a non-negligible depth of water and a continuous supply of water to the evaporating surface. To meet this goal, the meter consists of a vessel with a water volume of around 2 dl. The quick response and short-term accuracy are met by insertion of a micrometer into the lid of the closed, but ventilated, vessel. Exposure to insolation is met by using plexiglass. The whole set-up is considered as easy to read, maintain and rig up and it is also comparatively cheap. It can be used for standard readings once or twice a day but it can also be used for intensive studies where readings are taken every hour or more often. Because of its construction, readings have a strong variation with height. Andersson's (1969) evaporimeter is used primarily in Sweden, and to some degree in the other Nordic countries, but it is noteworthy that Fukuda and Suzuki (1964) present it as their own construction. It has been tested and compared with the class A pan and the Piche atmometer in Norway (Heldal 1969, 1973). Sundberg (1987) presents evaporation data, based on the Andersson evaporimeter, from 15 sites all over Sweden between 1975 and 1984. These data are intended to give information similar to that of the humidity maps presented by Eriksson (1986). Andersson's evaporimeter has also been used to estimate the atmospheric demand for water vapour in a joint Swedish-Tunisian study of shelterbed effects in central Tunisia (Åfors 1986; Paulsson 1987; Lewan 1987).

Johansson (1969) analyses Andersson's instrument in detail by constructing a model for its mass and energy balances. He also reports a thorough practical test of the meter. Readings of a few days were rejected because of increased water level in the meter after heavy rains, probable water loss because of storm winds, and ice in the meter. Johansson found a one-to-one correspondence between meter readings and Penman (1948) evaporation except in early spring and late autumn. As a result of his analysis he also presents a simple regression formula for potential evaporation

$$E_p = 0.14 + 3.70\ 10^{-3}\,Q + 0.13v(e_m - e) \tag{11}$$

Here E_p is in mm/day, Q daily insolation in cal/cm^2, v daily average wind speed at 1.5 m level (m/s), e_m saturated and e actual daily average vapour pressure at 1.5 m (mm Hg). Johansson's equation agrees well with that of Penman (1948) for summer periods but not at all for winter periods (Johansson 1970; Eriksson 1981). Johansson (1970) also compares his formula with those of Turc (1954, 1955), Thorntwaite (1948) and Blaney and Criddle (1950). It should be noted, though, that the validity of Eq. (11), for places other than mid Swedish farmland, has yet to be proven. Johansson (1973/74) uses the formula in a multireservoir model for the soil water balance. From the soil water distribution so achieved, Johansson calculates a set of multipliers for each reservoir and each phenological stage. The multipliers are finally used to transform the potential evaporation into actual evaporation. Johansson (1973/74) also finds a good correlation between the growth of barley under various fertilisation regimes, or productivity of pasture grounds, and the quotient of calculated actual evaporation to calculated potential evaporation (Eq. (11)) using data from Ultuna, mid Sweden.

Evaporation from Specific Plant Species

A study of wheat, barley and fallow evaporation at Ultuna, Uppsala, is reported by Sandsborg and Olofsson (1980). During the dry summers of 1975 and 1976, percolation could be disregarded and evaporation was determined as a soil water balance residual. Also the sensible heat flow was determined, using the energy balance together with accurate soil heat flow and radiometer readings (Rodskjer and Sandsborg 1980). From May to August those years an average of 130 mm per season fell as rain. For the same years, evaporation amounted to 300 mm per season for winter wheat and 175 mm per season for fallow. For barley, the corresponding figures from mid May to the end of August were 118 mm rain and 268 mm evaporation. It turns out that evaporation from fallow was fairly constant throughout the season, but evaporation increased with *LAI* (leaf area index) until midsummer for barley and wheat, after which it decreased until the stands were fully ripe. Around midsummer, the loss of sensible heat was negative, probably because of advective effects.

Linnér (1984) used Andersson's evaporimeter to estimate the potential evaporation from potato cultivations. He combines these data with water balance estimates of actual evaporation in order to establish a relation between potato yield and the quotient of actual evaporation over potental evaporation. Linnér reports values of potential evaporation between 260 mm and 340 mm from an experimental field at Ultuna, Uppsala, during the growth seasons (end of May to end of September) 1971-1975. The quotients between actual and potential evaporation varied seasonally with low values around 0.2 in early summer, values approaching unity for a long period in the middle of the growth season, and values of 0.4-0.6 for the autumn. Linnér (1984) also presents results from twenty field trials between 1975 and 1978 in southern Sweden, notably in Ugerup, where Högström and Larsson (1968) performed their studies in the early 1960s. Potential evaporation during the growth season varied between 304 mm and 418 mm, whereas precipitation for the same period was 56-182 mm. The average quotient between actual and potential evaporation varied from 0.44 for non-irrigated fields to 0.65 for well irrigated fields.

Lake Evaporation

Many lake evaporation studies were carried out in the beginning of the century. Studies of lake evaporation have been fairly neglected in recent years, perhaps because no major societal problem has demanded accurate data of this kind. The first investigation in modern time was carried out by a certain "major Nerman", whose data were used by Wallén (1914) in his evaporation methods review. Nerman measured evaporation from lake Hjälmaren from May to October between 1890 and 1899. He found average seasonal evaporation to be 408 mm during this period. Wallén (1914) uses these figures in combination with the Wild evaporimeter data from the observatory in Stockholm to find an annual lake evaporation of 580 mm. Further measurements from 1899 to 1914 at lake Hjälmaren are discussed by Wallén (1917), who presents weekly averages of evaporation for this period which are contrasted with data from the very dry year of 1914, when 640 mm evaporated during the growth season as opposed to the average value of 480 mm. Ångström (1920) critizes the use of pans to estimate lake evaporation. He used instead, for the first time in Sweden, an energy balance approach when he found evaporation from the northern Swedish lake Vassijaure to be 59 mm in August. Melin (1928) presents a thorough study of Lake Tåkern (44 km^2). With a water balance method he shows that this extremely shallow lake (average depth 1 m), with its small heat storage capacity, evaporated 760 mm per year. Melin (1928) used two different methods to estimate evaporation. For the period May to September he found the energy balance estimate to be 428 mm, whereas the water balance estimate was 601 mm. After refining some of the steps in the calculation, Wallén (1929) shows that this discrepancy is even larger than suggested by Melin (1928). Hjälmaren, one of the largest Swedish lakes, was investigated by A. Wallén (1934) with the water balance method and was found to yield an annual evaporation of 600 mm. Ångström (1939) found annual evaporation from the Motala River to be 480 mm by using data from a pan with artificially stirred water. Johnsson (1946) applied the energy balance approach to Lake Klämmingen. For the three years 1941-1943 he found an annual evaporation of 488 mm from the deep part of the lake and 505 mm from the shallow part (This includes estimated winter values added by the present author, which were personally communicated by T. Jutman 1982). The annual variation between the two parts showed much larger difference than did the annual total. The deep part of the lake evaporated less in spring and early summer, whereas the opposite held true for the autumn.

The IHD period brought an injection into lake evaporation research. Several investigations were carried out in the "representative" basins but most of them have not been published in the open literature. Rodhe (1973) was able to utilize radiation and lake temperature data, collected for other purposes, from the Velen experimental catchment. He estimates evaporation, with the energy balance method, to have been 353 mm between June and September 1971. This is contrasted to results from a class A pan (396 mm), a GGI 3000 pan (273 mm) and a water budget (385 mm). Rodhe (1977) discusses the possibilities to considerably simplify estimation of lake evaporation. The Bowen ratio can be expressed solely as a function of air temperature if the Priestly-Taylor (1972) equation for potential evaporation is combined with the energy balance. Since net radiation can be well predicted from global radiation through a simple regression, Rodhe (1977) proposes that lake evaporation can be accurately calculated from data of air temperature and insolation only. Data from three different lakes support the hypothesis but a definite approval of the theory requires further testing. It may be that increasing use of surface water may necessitate such tests in order to arrive at safer estimates on lake and stream evaporation.

Bringfelt (1988) uses the PHOENICS model to simulate possible differences in evaporation from an elongated lake for the two cases where wind is blowing either along the lake or perpendicular to it. He finds that the difference is small and presents, as a result of the simulations, his own parameter values for the bulk aerodynamic formula. Bringfelt (1988) then compares this formula with the Priestly-Taylor (1972) equation, and a modified Penman (1948) formula for lake Locknesjön in northwestern Sweden, for which water temperature data are available for the period 1953-1957. He shows that the accumulated evaporation during summer and autumn varies little, between years, from an average around 200 mm, and concludes that the Penman type equation is preferable since it is fairly insensitive to the corrections needed to transform climate data measured over land to representative data for a lake.

Snow Evaporation

Snow evaporation represents an area of low interest in Swedish evaporation research, except for a short period in the beginning of this century. Hamberg (1896b) reports snow evaporation rates at different air temperatures based on observations between 1888 and 1895. He used data from the previously mentioned Wild evaporimeter in Stockholm, which was filled with snow whenever available. He shows that maximum evaporation at a fixed temperature achieves a maximum of 2.0 mm per day at $-5\,°C$. At freezing, maximum is 1.4 mm per day, but at $-4\,°C$ it is 1.5 mm per day. Average daily snow evaporation continuously decreased from 0.56 mm at freezing, over 0.23 mm at $-10\,°C$ to 0.05 mm at $-20\,°C$. As a monthly average for January and February this gave just over 10 mm per month. Westman

(1901), who also used a Wild evaporimeter, found evaporation more frequent than condensation in Uppsala in March. Evaporation in this case did not exceed 0.6 mm per day and maximum condensation was 0.3 mm per day. Jansson and Westman (1902) present measurements from Uppsala with snow filled bowls during February to April. They found condensation to be negligible and evaporation to be less than 0.3 mm per 24 hours. Westman (1913) directly measured snow evaporation and condensation in southern Sweden using snow filled bowls. He found daily condensation less than 0.4 mm and daily evaporation less than 1.3 mm. Similar measurements carried out by Rolf (1914) in Vassijaure, northern Sweden, in a more humid climate during March and April indicate that condensation was even larger than evaporation. Rolf (1914) shows evaporation to be directly proportional to the vapour pressure gradient near the surface. Ångström (1918) computed daily evaporation using Rolf's formula, in Abisko, northern Sweden, during January, in order to calculate the energy balance at the snow surface. He found a net condensation for all day-time hours. The previously mentioned studies in the Lake Malmagen high mountain basin (Melin 1943) indicate a difference of 144 mm per year between evaporation and condensation that was not recorded in the precipitation gauges. C. C. Wallén (1948) made glaciological investigations at the Kårsa Glacier from 1942 through 1948. Wallén found that snow evaporated in May but that during summer only condensation took place.

A sophisticated lysimeter was devised by Nyberg (1966) to find out the net effect of snow evaporation and condensation. Between 8 March and 7 May 1962 he measured 27 mm evaporation and 5 mm condensation in northern Sweden. He also gives a set of regression equations for condensation and evaporation based on vapour pressure gradient and wind speed. A simpler snow lysimeter, devised by Jansson *et al.* (1978), consists of an inner bucket with a perforated bottom, placed in a slightly larger outer bucket. The whole set is to be placed in level with the snow cover. Measurements with this lysimeter are reported by Bengtsson (1980), who also refers to water balance studies carried out at the Lappträsket representative basin. These show that snow evaporation over the whole winter only amounts to some few mm. As a conclusion, Bengtsson states that snow evaporation is only of minor importance in the annual water balance. For northern regions it normally amounts to no more than 10 to 20 mm per season. From an energy point of view, however, snow evaporation is an important factor which interacts strongly with snow melt. Since no conventional formula of snow evaporation accounts for the actual heat exchange processes above a snow surface, this brought Bengtsson (1981) to construct a more physically based model founded on Monin-Obukhov similarity theory. The model accounts for atmospheric stability and melting or freezing at the surface. Output from the model is surface temperature, melt rate, sensible heat flux and evaporation/condensation.

Bringfelt (1985b) attacks the problem of evaporation from a heterogeneous snow cover with patches of bare ground covering a smaller or larger part of the area. He uses the PHOENICS model to simulate the net evaporation from a surface which is alternately covered by snow and wet ground. In such cases, there is normally condensation at the snow surface and evaporation from the ground, and in most situations the net result is an evaporative flux into the atmosphere. On the basis of the simulations, Bringfelt (1985b) presents a set of formulæ to calculate snow evaporation/condensation for all situations ranging from full snow cover to totally bare ground. This set is used to calculate monthly evaporation/condensation during March to July from 1973 to 1985 at Katterjåkk and Storlien in northern Sweden. Bringfelt (1985b) presents his data as a range depending on the assumed bare ground temperature. His calculated maximum accumulated evaporation for the period March-June is 33 mm for Katterjåkk in 1982, whereas his minimal value for the same period corresponds to a condensation of 9 mm at Katterjåkk in 1973.

Discussion

Stanhill (1973) investigated the amount and nature of the international evaporation research literature. He found that there had been an exponential growth of such literature at least since the 1950s such that the publication rate was several hundreds of papers annually in the beginning of the 1970s. Stanhill states that it is impossible, under such circumstances, to know what has been published in this field. This means that a considerable degree of replication is inevitable. Since approximately every second publication only presents results of water loss measurements, he questions the justification of their continued appearance in the scientific literature. Whether one agrees or not with Stanhill's conclusion, there remains the problem of valuable results not being published. As the previous paragraphs have shown, the IHD programme meant a vital injection into Swedish evaporation research, but valuable results from the IHD period still reside in the drawers of many researchers. This may not be so serious in well studied areas but the effect is detrimental when it comes to the understanding of, *e.g.*, bog evaporation. There is, of course, also in Swedish literature an amount of less valuable reiterations. During the preparation of this review I have, for instance, come across several, fairly thick, reports dealing with computer programming of Penmans's equation. Some of them deal with simple sensitivity tests and some present preliminary results of simple applications, but most of them present the results as being scientific breakthroughs.

Many problems in, *e.g.*, hydrology or agriculture have been solved sufficiently well with evaporation data given as long-term residuals or calculated with some standard empirical formula. This is somewhat curious since earlier maps of evaporation have been systematically erroneous by almost 30 % and since modern research has shown that the most accepted formula, that of Penman (1948 or 1953), is not applicable to evaporation from the major part of Sweden's land area. There are, in addition, several problems, specific to the present time, that cannot be solved by the old type of large-scale and long-term residual evaporation estimates. Such questions relate to changes in land use, planning of irrigation, dimensioning of biological filters to prevent leaching from refuse dumps, interpretation of remote sensing data, input to meteorological models, improvement of the water use effciency of different plant species, etc. A few questions also relate to classical hydrology. One of them has to do with the possibilities to explain a lacking balance between runoff and precipitation input in winter with a non-negligible evaporation of intercepted snow. Another has to do with the possibilities to explain a long-term trend of decreasing runoff that might be caused by the increased areal extent and productivity of Swedish forests (personal communication, B. Eriksson 1988). The answer to all these questions demands a knowledge of evaporation with a high resolution in time and space, as well as of the structural properties of the evaporating surface.

Estimates of evaporation can differ widely between different investigators and at different times. It has been shown that forest evaporation is assumed to vary between 250 mm and 800 mm annually. This wide range is a bit confusing since the data all emanate from the same climatic region. There are several explanations to this situation.

Firstly that measurement of evaporation seems to be a very delicate business. It has been shown in several investigations that quite different results can be achieved if more than one measurement method is used. The most remarkable of these discrepancies is reported by Grip *et al.* (1979) and Lindroth (1984b) who show two independent measurements to differ by more than 100 %. It is suggested by Grip *et al.* (1979) that this might be explained by a lack of validity of the similarity assumption in the Bowen ratio theory. Continued measurements have also shown that there are basic problems associated with the traditional assumptions of flux-profile relations above the Jädraås forest (personal communication, U. Högström 1987). Since many traditional methods have been shown to fail, there is an obvious need to develop new or to refine old methods for evaporation measurements. More emphasis should be placed on the only direct micrometeorological method, *i.e.*, the eddy correlation technique, and the possibilities to develop this for further use by the research community. Since it is not realistic to believe that this technique will be applicable on a routine basis, simplified methods are urgently needed. New ideas should be encouraged, like when Rodhe (1987 page 140) suggests that snow evaporation might be calculated from isotopic fractionation.

Secondly that earlier measurements and estimations are based on a theory of exchange processes between a vegetative surface and the atmosphere that has been proven erroneous or insufficient. There are several aerodynamical aspects of a vegetation cover that defy the traditional exchange theory, the most eye-catching being the regular appearance of countergradient fluxes (*e.g.*, Denmead and Bradley 1985). Halldin and Lindroth (1986) try to apply the traditional flux-gradient theory for exchange within the Jädraås forest canopy, but conclude that further progress will require development of more sophisticated models. Such models are proposed by, *e.g.*, Wilson and Shaw (1977) and Raupach (1987). The development of these models are hampered by the absence of field data on turbulence properties within the canopy and in the roughness sublayer immediately above it.

Thirdly that previous measurements and estimates have been based on more or less restricted data sets. It has been shown, at least for forest canopies, that a correct description of stand evaporation requires that transpiration and interception evaporation be treated separately, and also that the surface resistance of the stand be emphasized. It will be necessary to encourage development, testing, and refinement of models that are physically realistic, yet no more complicated than what is needed for their purposes or permitted by the availability of structural input data.

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