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Using rainfall-runoff modeling to interpret lake level data *

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Abstract

Using water balance computations, the behavior of different kinds of lakes is discussed. Simple analytical expressions relating water level to hydrological conditions and lake bathymetry are given. The importance of knowing the river basin area when analyzing lake levels is stressed. A conceptual rainfall-runoff model including lake routing is used to simulate runoff and lake levels and to compute quasi-steady state conditions and long-term transient situations. It is suggested that models can be used to construct curves relating lake levels to precipitation and lake evaporation. By comparing with paleo-lake levels, the annual precipitation related to these levels can be found, provided information is available about the seasonal distribution of the precipitation.

Introduction

Paleolimnological studies can give information about previous lake levels, the productivity and the temperature of the lakes. The conclusions to be drawn from such information about previous hydrological and climatic conditions are not obvious. Changed lake levels may be due to changed precipitation and/or changed evaporation, to changed conditions at the lake outlet or downstream of the lake, or due to changed conditions in the upstream river basin. Reconstructed lake levels have been used to infer changes in the hydrological balance, e.g. Street and Grave (1976), and also to infer paleoclimatic changes at large scale (Street-Perrott & Harrisson, 1985). As pointed out by Harrisson (1988) much more information about hydrological conditions is obtained from closed basin lake levels than from lakes which overflow, since the open lake can not rise very high because of related high outflows. A lake may change status to go from an open lake to a closed lake when the climate becomes dry, and then back again. Lake Mendota, Wisconsin is such an example as shown by Winkler (1985) cited by Harrisson (1988). Often, reasons for lake level changes can be given from simple hydrological considerations, e.g. Belyaev (1991). Also, the prerequisites for lake level changes to occur in different types of lakes can be found from rather simple considerations. However, since the basin runoff response to precipitation is highly non-linear, mathematical rainfall-runoff models and lake routing models must be used in order to quantify the lake level response to different climatological conditions.

In this paper, after a short review of different types of lakes has been given, it is shown how lakes of different character generally and in principle respond to variations in precipitation and evaporation, in a short- and long-term perspective. It is discussed during which conditions different types of lakes can exist and how long time is required for the lakes to adjust to new climatological conditions. A rainfall-runoff model which includes a detailed lake routing procedure is used to compute lake level variations in detail. It is thought that by comparing computed and reconstructed levels, conclusions can be drawn about previous precipitation and evaporation.

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Different types of lakes

Most lakes are open lakes, i.e. they flow over a sill into an outflowing downstream river. The inflow is for most lakes streams and rivers. There may be periods of ceasing inflow. The water level may then drop to near, or even below, the outlet threshold, but within a couple of months the lake again flows over. When the inflow is large, the lake level rises, which results in high outflow. Since in this way high inflow is compensated for by high outflow, the lake level remains at a moderate level not very much above the sill level. In lakes with river inflow, the groundwater contribution is usually very minor. A lake situated in a large aquifer may be groundwater spring-fed. The spring occupies a restricted area in the deep part of the lake, and the inflow is usually large relative the lake area. In eskers or in areas of limestone, the groundwater effluent seepage may be diffuse around the shores of the lake. In Scandinavia, where there are only few groundwater effluent lakes, these lakes are situated just downstream eskers and constitute themselves a significant part of their river basins.

In semi-arid climate, lakes may be terminus lakes, with inflow but no outflow. If such closed lakes are situated above the sea level, there may be some seepage so that the lake at times looses minor amounts of water to groundwater. Net groundwater outflow has for example been reported by Richardson & Richardson (1972) for Lake Naivasha in Africa.

Lakes situated very high up in a river basin, at or close to the water divide, are mainly controlled by atmospheric exchange of water. Anderson & Munter (1981) introduced the concept of recharge lakes and closed lakes for lakes having no input except precipitation. A closed lake is cut off from the groundwater and closed from connections with other surface waters. With this definition a closed lake can only exist if there is a perfect balance between precipitation and lake evaporation. A recharge lake also receives precipitation as the only input, but has according to the definition by Anderson and Munter groundwater outflow. It can exist only if the precipitation exceeds the lake evaporation. With a more general definition, recharge lakes include all lakes without surface or groundwater inflow and thus many high raised bog-pond-small lake systems, which are so common in Scandinavia and northern Canada.

In a water balance classification system worked out by Szestay (1974) and further developed by Street (1980) and Street-Perrott and Harrison (1985), the concept of atmosphere-controlled lakes was introduced. These are lakes for which the precipitation on the lake considerably exceeds the river inflow and the lake evaporation exceeds the outflow. These lakes are in principle recharge lakes although they may have a small upstream river basin.

Observed lake levels

Many of the Swedish data series of daily lake level observations go back to the early decades of this century (Lindkvist & Danielsson, 1987). There are even some series starting in the early 1800s. Some lake level series are shown in Figure 1. It is seen that the seasonal lake level fluctuations are much larger than the differences of the annual mean level between different years. In Lake Siljan the maximum mean annual water level for the period is 1.7 m above the outlet sill and the minimum 1.1 m, which is a range of only 0.6 m, while the range between maximum and minimum daily water levels is almost 4 m. Historical floodmarks (Harlin, 1989) show that the lake rose to about 6.5 m above the outlet sill in 1659 and in 1764. Such occasions may be due to combinations of intense melt and rainfall, but may also be caused by blocking of the outlet for example by ice jamming, in which case the high level is very temporary.

Lake Siljan is 354 km^2 and the lake area constitutes 3% of the river basin. The much smaller Lake Möckeln, 45 km^2, constitutes 4.5% of its river basin area. The outlet is narrow, so the water level is quite high above the outlet sill. The mean annual water level, for the period shown in the figure, varies between 1.6 and 2.1 m. The seasonal water level fluctuations in Lake Möckeln are of the order 1 m, with a maximum range of 1.6 m. The reason for the larger seasonal fluctuations in Lake Siljan compared to Lake Möckeln is because of the much larger amounts of snow accumulated in the river basin upstream Lake Siljan.

In the large, 1900 km^2, Lake Vättern with such a large ratio lake area to river basin area as 0.30, which means that the lake is not far from being an atmospheric controlled lake, the maximum seasonal range during the period shown in Figure 1 is 0.35 m, and the difference between maximum and minimum annual mean lake level is 0.2 m. From 80 years of data before the lake was regulated, it can be found that the maximum recorded lake level is 0.5 m and the minimum 0.05 m above the outlet sill. Generally, the larger the lake is

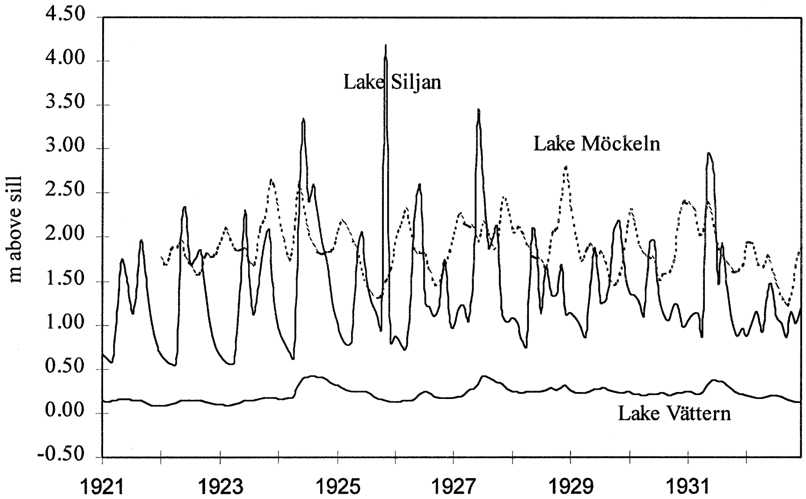

Figure 1. Observed water levels in Swedish lakes, 1921–1932.

relative to its river basin, the smaller the lake level fluctuations are.

Intuitively there is a relation between precipitation and lake level. However, since river inflow to a lake is dependent also on evaporation and there is a delay between precipitation and river flow, the relation is indirect. Still, when the mean lake level in Lake Möckeln is compared with the annual precipitation (Nov–Nov to avoid the direct influence of snow precipitation and snow-melt) as is done in Figure 2, it is seen that there is a close relation. This shows that lake levels adjust fast to changing climatic conditions, when the lake level is above the outlet threshold.

Lake levels depend on the lake level-outflow conditions. However, whatever the outlet conditions are, high water levels are counterbalanced by high outflows. To see the effect of the outlet conditions, lake levels were computed from lake level-outflow relations for Lake Siljan and Lake Möckeln for different hypothetical outlet conditions being given the water balances of the lakes. It was assumed that the outlets of the two lakes were widened to twice their present width and also reduced to half their present width. When widening the outlet of Lake Siljan, the mean lake level should drop from 1.4 to 0.9 m above the sill, and when reducing the outlet width rise to 2.2 m. The mean level of Lake Möckeln, today almost 2 m above the outlet threshold, would change 0.4 m up or down if the outlet width were to be reduced by 50% or increased by 100%. Thus, even rather drastic changes of lake outlet conditions do hardly influence lake levels to the extent that it can be quantified by paleolimnological methods.

Some basic and rather intuitive conclusions about overflowing lakes were drawn from the above section about observed levels in three Swedish lakes. First, seasonal lake level variations exceed by far interannual lake level variations. Second, the larger the lake is relative to its river basin, the more stable the lake levels are. Third, when the annual precipitation changes, the lake level adjusts fast to the climatological conditions. Fourth, major changes of lake outlet conditions only marginally influence mean lake levels.

Water balance of overflowing lakes

The water level of a lake depends on a balance between inflows and outflows. The inflows are precipitation on the lake, river inflow, groundwater inflow and diffuse overland flow. The outflows are river outflow over an outlet sill, groundwater flow and evaporation from the lake. The groundwater exchange is in most situations into the lake, and is, except in carstic areas or if the lake is spring-fed, almost always negligible compared to the other flows. Overland flow is minor. The water balance of a lake is

$$Q_{\text{in}} - Q_{\text{out}} + (p - pe)A = A\partial h/\partial t \qquad (1)$$

where in- and outflows are denoted Q_{in}, Q_{out} and can in principle be surface flow as well as groundwater flow, p and pe are the atmospheric mass exchange with the

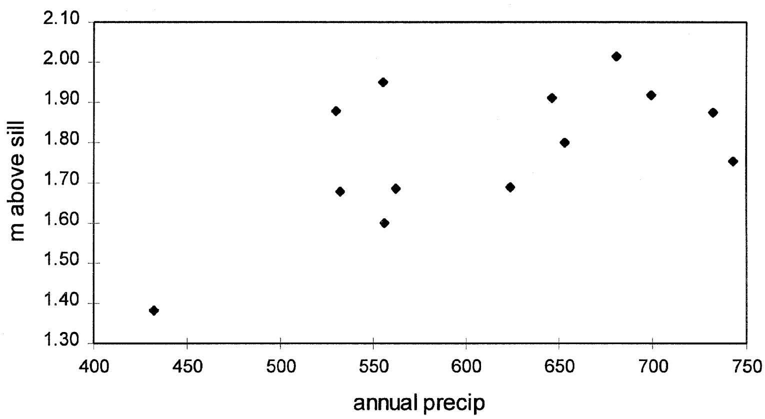

Figure 2. Mean annual Lake Möckeln water levels (1 Nov–1 Nov) versus the precipitation of the same periods.

lake, i.e. precipitation and lake evaporation, h is lake water level, t is time and A is the lake surface area, which depends on the lake level. The lake evaporation is assumed to correspond to the potential evaporation.

If the lake is not affected by damming from downstream, there is a unique relation between outflow and lake level. Knowing this stage-discharge relation $Q_{out}(h)$, which if the groundwater outflow is negligible can approximately be put in the form

$$Q_{\mathrm{out}} = b(h - h_{\mathrm{sill}})^m \qquad (2)$$

where h_{sill} is the outlet sill level, and b and m are coefficients which depend on the river outlet conditions, and knowing the bathymetry of the lake A(h), daily lake level variations and long-term quasi steady state levels can be determined.

Using q to denote specific runoff, the river input to a lake is $Q_{in} = q\ (A_b - A)$ where A_b denotes the river basin area of a lake including the lake itself. Substituting Q_{in} in the water balance Eq.(1) by the expression above and Q_{out} by Eq.(2), a steady-state lake level is found as

$$h - h_{\mathrm{sill}} = \left(\frac{(p - pe)A + q(A_b - A)}{b} \right)^{1/\mathrm{m}} \qquad (3)$$

It is seen that if the lake constitutes only a small fraction of the total river basin and if more than very little runoff is produced, the precipitation on the lake and the lake evaporation are of little importance for the lake level.

The coefficients of the stage-discharge equation depend on the river outlet conditions, i.e. the outlet river cross section, and the bottom slope and the bottom roughness of the outflowing river. As already shown, lake level variations depend on the outlet conditions only in the range of half a meter or so. To illustrate also the effect of changing hydrological conditions, i.e. different annual precipitation, evaporation and runoff, in a lake with river outflow, lake levels were computed from Eq. (3) for different p, pe, q and different ratios lake area to river basin area. Rather arbitrary, the outflowing river was assumed to have a bottom slope of $S = 10^{-3}$, a Manning roughness of $n = 0.05$, which is representative of a small stream with graveled bottom, and a triangular section with shores sloping mildly at $\alpha = 15\ °$. When the river flow is uniform with depth h_r so that the Manning formula ($Q_{out} = S^{1/2}/n\ A_w{}^{5/3}\ P_w{}^{2/3}$) can be applied, the outflow coefficients are b = 1.5 and m = 8/3, since the wetted area is $A_w = h_r{}^2/\tan\alpha$ and the wetted perimeter, i.e. the bottom part of the cross section along which there is friction, is $P_w = 2h_r/\sin\alpha$ and therefore b is

Table 1. Mean annual water level above the outlet sill in an overflowing lake for different meteorological (potential evaporation minus annual precipitation, pe-p), hydrological (annual specific runoff, q) and physiographical conditions (lake area fraction of the river basin including the lake itself, A / A_b). The outflowing river is assumed to have bottom slope 10^{-3}, a Manning roughness 0.05 and to have a triangular cross-section with shores sloping 15°.

Lake area fraction		0.001	0.01	0.10
pe-p (mm)	q (mm)	level (m)	level (m)	level (m)
0	200	1.7	1.7	1.7
500	200	1.7	1.7	1.5
0	100	1.3	1.3	1.3
500	100	1.3	1.3	0.9

$(\cos\alpha)/(2^{2/3}\sin\alpha)$. The results of the computations are shown in Table 1. It is seen that even drastic climatic changes only have minor effects on lake levels as long as the lakes overflow, which is due to the dependence of lake level on river outflow. The lake must constitute at least 10% of the river basin for the net atmosphere-lake mass exchange, pe-p, to influence the lake level, since otherwise the river basin runoff completely dominates the atmosphere-lake mass exchange. Further, this mass exchange must be a loss to the atmosphere, i.e. pe>p. It is only when the climate is drastically changed, for example (pe-p) increased by 500 mm yr^{-1} and the annual runoff decreased by 100 mm, that the annual mean lake level is considerably changed, in the example reduced from 1.7 m to 0.7 m above the sill, and then, as seen from Table 1, only for a lake which constitutes a rather large fraction of a river basin, in the example 10%.

Water balance of lakes with no outflow

When the specific runoff is very low and the evaporation loss from the lake is high, the water level may drop below the sill, first only temporally and then for the whole year. It is easy to see from water balance calculations, Eq.(1), and, for steady-state conditions, in principle the nominator inside the parenthesis of Eq.(3), that for a lake not to overflow, the lake area at sill level must be larger than the upstream river basin area times q/(pe-p). For quasi steady state conditions to prevail, the lake surface area must be a certain fraction of the whole river basin area,

$$A/A_b = \frac{q}{q + pe - p} \tag{4}$$

In some paleolimnological literature, the ratio lake area to upstream river basin area excluding the lake itself is used and called z-value (Mifflin & Wheat, 1979; Bowler, 1981). However, there is an advantage in using the ratio lake area to whole river basin area, since this ratio is restricted to values between zero and unity, whereas the z-ratio can approach infinity. Mifflin and Wheat give z-values for lakes in the Great Basin in SW USA. In the more humid parts near the western mountains the z-values corresponds to ratios of A/A_b in the range 0.4–0.6, and in the desert area in the east to 0.04–0.3, which manifests that the lake area must constitute a large part of its river basin not to spill over, unless the climate is extremely dry.

Obviously, for a lake not to have any outflow, the lake evaporation must exceed the precipitation on the lake, but also the specific runoff must be quite low. Some required combinations of net atmospheric water loss (pe-p), specific runoff and lake area to river basin area ratios for a lake not to have any outflow are shown in Figure 3. In a climate where the annual lake evaporation is not very much larger than the annual precipitation and the runoff not very small as for example in southern Sweden, p = 600 mm, pe = 700 mm, q = 200 mm, the lake must constitute at least 2/3 of the total river basin for the lake not to spill-over. In a dry hot climate, p = 200 mm, pe = 1200 mm, q = 20 mm, it is enough if the lake area is 2% of the river basin, but even that is a lake percentage higher than observed in most river basins.

If the lake level permanently is below the outlet threshold, the lake level, or rather the lake area, adjusts itself so that the net evaporation loss to the atmosphere balances the river inflow. The bathymetry of a lake can be expressed as a function of the lake level, A = f(h). For a lake with uniformly sloping shores the surface lake area is related to the lake depth as

$$A = ah^2 \tag{5}$$

which, when A is substituted in Eq.(4) gives the lake depth

$$h = \sqrt{\frac{A_b}{a}} \cdot \sqrt{\frac{q}{q + pe - p}} \tag{6}$$

The ratio between equilibrium depth for different climatic and thus different hydrological conditions is

$$\frac{h_1}{h_2} = \sqrt{\frac{q_1}{q_2}} \cdot \sqrt{\frac{q_2 + pe_2 - p_2}{q_1 + pe_1 - p_1}} \tag{7}$$

where index 1 denotes a certain condition and index 2 another. When the lake area is a general function of the water depth, A = f(h), the square root of Eq.(6) should be replaced by the inverse function h =
f^{-1} [- A_b q/(q + pe-p)].

For example, if a dry hot climate, p = 200 mm, pe = 1200 mm and q = 20 mm per year changes into a less dry climate, p = 500, pe = 1000 and q = 100 mm, the new water depth in a lake with uniformly sloping shores is almost 3 times the old water depth. Clearly, because there is no outflow, the level of a terminus lake is much more sensitive to hydrological changes than a lake which overflows.

In a circular lake with shores sloping 1:s, the coefficient a in Eq.(5) is Π s^2. This means that if the river

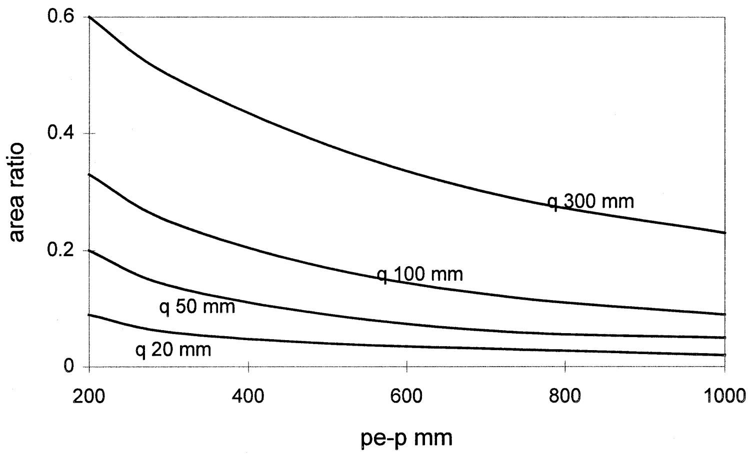

Figure 3. Minimum ratio lake surface area at sill level to total river basin area (A/A_b) for a lake not to have any outflow for different climatological conditions (potential evaporation minus annual precipitation, pe-p) and different hydrological conditions (annual runoff, q).

basin area is 100 km^2 and the lake shores are sloping 1:50, the lake depth of the dry hot climate conditions should be 15 m and of the less hot climate 45 m. Also, the lake area increases very much from, 2 km^2 to 17 km^2. For a lake not to overflow when the climate is not extremely dry and yet not to dry out when the climate is very dry, the lake must be deep. Indeed, Street-Perrott and Harrison (1985) referring to Smith (1979) show that the Searles Lake today has a size of only 100 km^2 but once covered an area of 1000 km^2 and had a depth of 200 m.

The world's largest lake or inland sea the Caspian Sea is chosen to illustrate the use of the steady-state closed lake equation (4). Vali-Khodjeini (1991) gives estimates of river inflow, lake evaporation and precipitation on the lake. The precipitation over the lake is highly spatially variable in the range 100–1700 mm per year. The annual lake value seems to have varied in the range 180–240 mm in this century. Also the lake evaporation varies. From a Russian literature review by Belyaev (1991) it seems that the annual value is 995 mm. The highest and lowest estimated annual values do not deviate more than 10% from this value. The mean inflow in the early part of the century corresponded to a specific runoff of 90 mm per year. Using a high precipitation value and a low annual evaporation, the lake area is using Eq(4) computed to be 11% of the total river basin. The lake area in 1930 was about 405 000 km^2, which is 10% of the river basin. The inflow in the period 1937–1977 corresponded to an annual runoff of 73 mm. Using low precipitation and high evaporation as input in Eq.(4), the lake area for quasi steady-state conditions should be 8%. The lake area constituted in 1977 about 9% of the river basin. Thus, a steady state condition was not yet reached. Instead, the Caspian Sea has risen as an effect of increased precipitation in the river basin and consequent increased runoff.

Seasonal and interannual lake levels

When observed levels in overflowing Swedish lakes were discussed, it was found that the mean levels do not change much from year to year, but the seasonal variations could be a couple of meters. From the discussions on water balance it was found that the mean water levels in lakes with outflow do not change very much when the climate is changed, but that the water level of lakes without outflow drops and rises considerably if the climate becomes dryer or wetter.

To see the lake level variations over longer time periods, the runoff from a hypothetical but typical southern Swedish river basin and the water level in a small downstream lake were simulated using a conceptual runoff model, which is briefly described below, with meteorological data from Lund, latitude N56 °, over the period 1753–1990 as input. The annual precipitation, about 600 mm, for the period are shown in Figure 4. The computed annual mean and maximum water levels are shown in Figure 5. The typical seasonal lake level variations are shown for the year 1800 in Figure 6. The lake area is 1 km^2 and the river basin including the lake is 10 km^2. The outlet from the lake is assumed to be a river rapid with a 90 ° triangular cross section, i.e. the outlet constitutes a hydraulic control section. This means that $Q_{out} = 1.1\ (h\text{-}h_{sill})^{2.5}$ in SI-units with $(h\text{-}h_{sill})$ as the lake level above the lake outlet threshold. The potential evaporation is calculated from the air temperature using the Thornthwaite (1948) formula. The annual potential evaporation is about 700 mm. The precipitation varies in the range 300–800 mm yr^{-1}, but the computed annual mean lake level only between 15 and 40 cm above the outlet threshold. The computed water level did never on any single day exceed 1 m above the sill. It is seen from a close examination of the curves that there is a rather close relation between precipitation and annual mean water level, which was also found from the study of Lake Möckeln. The simulations showed that the very low 300 mm precipitation in 1868 with a computed lake level of 15 cm above the outlet threshold followed by a year of 650 mm precipitation, resulted in an adjustment of the lake level to 30 cm above the sill already the same year.

All the simulations in this paper are performed using a rainfall-runoff conceptual model called HBR (Bengtsson et al., 1995), which also includes the lake routing procedure used when lake levels and lake outflow are computed. The HBR-model is a conventional conceptual model of which there are at least 50 similar ones around the world. The hydrological part of the model is rather conventional with solid snow and liquid in snow storage, soil water storage, groundwater and separate deep groundwater storage. Runoff can occur as overland flow, interflow, quick groundwater flow, slow groundwater flow and deep groundwater flow. As in all conceptual models, for each storage box, continuity and outflow-storage relations are used to compute the outflows and the state of each storage at each time step. HBR was run with daily precipitation and air temperature as input, but with much shorter computational time step. In the applications in this paper, the runoff was dominated by slow and quick groundwater flow. The model can be distributed into many sub-basins connected by rivers, and each sub-basin can be divided into many zones mainly to distinguish between different land use and soil conditions. However, for the simulations in this paper no division into sub-basins was done. The hydraulic part of the model is lake and river routing of water through a large river basin. In the present study the lake routing procedure was applied to a lake downstream a river basin. The lake routing is based on the lake water balance Eq.(1) and relations between outflow and lake level, as for example Eq.(2) although more complex relations can be used, and between lake surface area and lake level, as for example Eq.(4) although again much more complex relations are commonly used.

There are some 30 parameters in the model, but since lakes even out river flows, only 2–3 parameters significantly influence the lake level computations. The annual river basin runoff, and thus also the outflow from a lake, is, disregarding some minor interannual storage, simply the difference between annual precipitation and basin evaporation. Thus, if the annual basin evaporation is modeled correctly, also the outflow from a downstream lake is modeled correctly and, because of the relation between outflow and lake level, also the lake level is modeled correctly. In the model simulations, parameters typical for forested areas in southern Sweden were used. For computing seasonal runoff the model is quite robust giving accurate output even when the model parameters are not chosen with great care. For computing lake levels the model is even less sensitive to the choice of model parameters.

After having simulated lake levels in an existing climate, a 250 year period, the HBR rainfall-runoff model simulations were repeated for changed climatic input. The new meteorological input data were the previous observations from Lund manipulated to give different annual precipitation and potential evaporation, still keeping the relative seasonal precipitation, potential evaporation and temperature distributions, and also not changing the daily precipitation and temperature distributions within a month, unless stated otherwise in the paper. This means that observed daily precipitation were multiplied with given factors, and observed air temperature changed by a given value.

The computed specific runoff for different scenarios of combinations of precipitation and potential evaporation not changing the seasonal distribution of today is given in Figure 7. The runoff decreases to very low values when the annual precipitation falls below 250-300 mm. Since basin evaporation can not take place at the potential rate during summer, variations in potential evaporation have less influence on runoff than precipitation variations have.

In the existing climate of southern Sweden with lake evaporation barely exceeding the annual precipitation, closed lakes without outflow can exist only, if the upstream river basin excluding the lake is smaller than the lake itself. However, if the annual precipitation prevails below 250–300 mm, the previously discussed hypothetical lake, the lake level of which was simulated for the period 1753–1990 and which constitutes

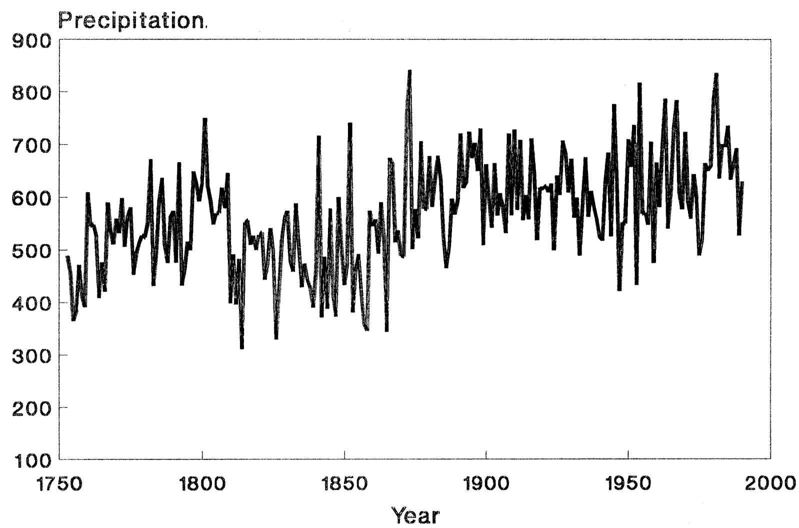

Figure 4. Observed annual precipitation, Lund, Sweden.

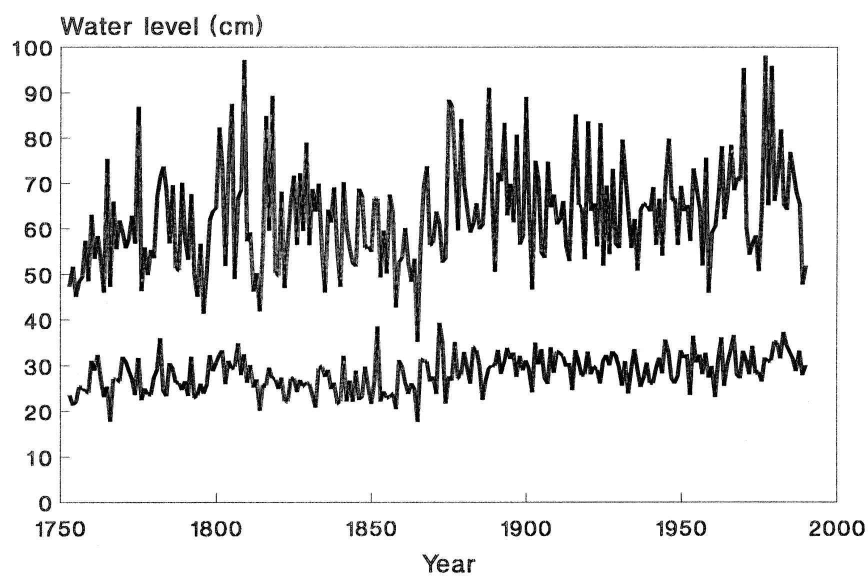

Figure 5. Computed annual mean (lower curve) and maximum (upper curve) water levels for a lake constituting 10% of a 10 km^2 river basin, when the lake stage – outflow relation is 1.1 $Q^{2.5}$, as determined from rainfall-runoff modeling and lake routing, using meteorological data from Lund.

10% of a river basin, ceases to spill over. Then, the lake level is sensitive to annual variations of precipitation. Conceptual runoff simulations followed by lake routing, which in this case, since the water level is below the sill, are simple water balance computations, show that, if the lake has steep shores, the water level would drop by 60 cm a year when the precipitation is only 50 mm, and increase by almost 1 m, unless there is outflow, in a year when the precipitation is 500 mm.

The seasonal distribution of precipitation influences the river basin runoff and thus the water level in a downstream lake. The annual runoff is higher when the rain falls in the cold period of the year, or when there is snowmelt, as compared to when most of the rain falls in periods of high potential evaporation. To show the influence of the seasonal distribution of the annual precipitation on the annual basin runoff and thus on mean lake levels, simulations were done

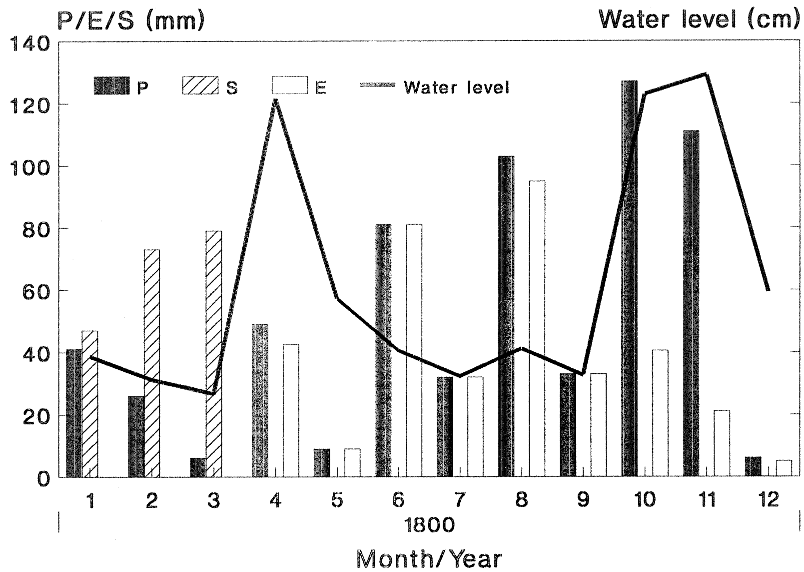

Figure 6. Observed monthly precipitation, P, computed snow accumulation, S, basin evaporation, E, and lake level for the year 1800.

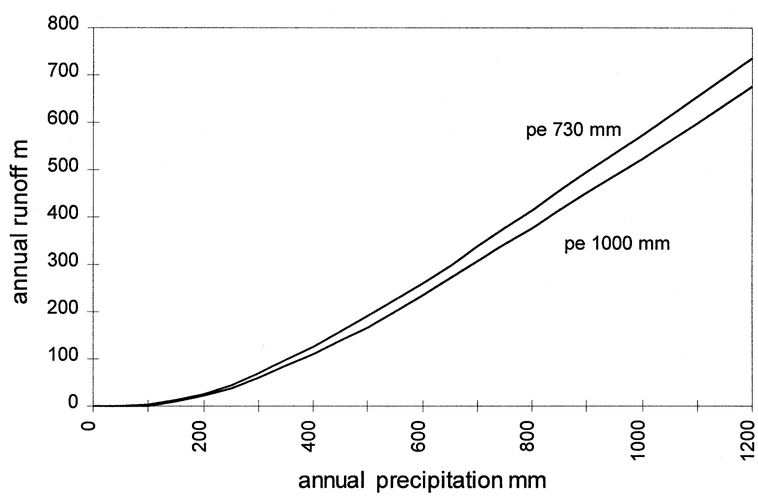

Figure 7. Simulated annual specific runoff from a forested South-Swedish river basin as function of annual precipitation for two different annual potential evaporation (pe) situations assuming the relative distributions within the year of precipitation and potential evaporation being the ones of the present climate.

for constant annual precipitation, 550 mm, and constant annual potential evaporation, 730 mm, but with different seasonal precipitation distributions. The precipitation was distributed as snow and rain, no snow but rain in summer as well as in winter, summer rain only, and winter rain only. The computed runoff and lake levels are shown in Table 2. Lake levels were computed for a lake with an area of 0.3 km^2. The basin area is 100 km^2, which means that the lake is 0.3% of the river basin. The outflow-lake stage relation is the same one as before assuming the outlet being a river rapid, $Q = 1.1\ h^{2.5}$.

The winter rains result in much higher annual runoff than a precipitation distribution dominated by summer rains, 310 mm as compared to 110 mm. However, although the runoff is very dependent on the precipitation distribution, the lake level does not, in spite of the lake area being such a small fraction of the river basin, change much, the mean level remaining at about 0.25 m above the sill in both scenarios. It can

Table 2. Annual specific basin runoff (mm) and lake level above outlet threshold (m) (triangular shaped outlet to a river rapid), computed for different seasonal distributions of snow and rain. Annual precipitation 550 mm. Potential evaporation 730 mm. Lake area at sill level 0.3% of river basin area. When precipitation was simulated to occur in winter as well as in summer (the three first scenarios), the seasonal precipitation distribution of the South-Swedish climate of today was used

	Runoff	Max lake level	Min lake level
As today, moderate snow melt	160	0.41	0.13
As today, intense snow melt	190	0.61	0.11
As today, no snow - only rain	150	0.37	0.16
Only rain and only in winter	310	0.62	−0.09
Only precipitation in summer	110	0.39	0.09

be noted that if all the rain falls in the winter, the lake level drops below the sill in the summer, although the annual runoff is the highest for this precipitation distribution. This example shows that there is no simple relation between annual precipitation and lake level in lakes with outflow. Thus, to be able to estimate annual precipitation from lake levels, it is necessary to have some information or make some assumption about the seasonal distribution of the precipitation.

The water level in closed lakes is as seen for example from Eq.(6) related to the annual runoff. Since in turn the annual runoff is strongly related to the seasonal precipitation distribution, the seasonal precipitation distribution strongly influences the lake level in lakes without outflow. For a lake of uniformly sloping shores, Eq.(7) gives, when the runoff from Table 2 and the precipitation and potential evaporation used when computing this runoff are inserted, the ratio between lake levels in a climate with only winter rain and in a climate with only summer rain as 2.2. Thus, for the same lake, the water level in a climate with rain in the winter only and no rain in the summer would be 20 m, when in a climate with summer rains only the mean lake level is 9 m.

Transition to quasi steady-state condition

When the climate is changed but a lake still is an open lake with outflow, the lake level adjusts fast, within a year, to the new conditions, as shown in previous sections. In a lake without outflow the adjustment takes much longer time. The transient condition of a lake without outflow approaching an equilibrium steady-state condition can be described by the water balance Eq.(1). The climatic conditions are changed to new pe, p and q, from which values the equilibrium lake area, A_∞, is given by Eq.(4). When dimensionless area, $x = A/A_\infty$, and dimensionless depth, $y = h/h_\infty$, h_∞ being the equilibrium depth corresponding to the lake area A_∞, is introduced, recalling that there is no outflow and using Eq.(4), Eq.(1) is after division by A_∞

$$1 - x = \frac{h_\infty}{(pe + q - p)} x \frac{\delta y}{\delta t} \tag{8}$$

Also a dimensionless time, $\tau = t/\tau_{scale}$, can be introduced if a time scale, τ_{scale}, is defined. From dimension analysis it is clear that the time scale has to be the lake depth divided by either evaporation, precipitation or runoff, or some combination of the three. It is seen from Eq.(8) above that a time scale related to the transient process is

$$\tau_{scale} = \frac{h_i nfty}{q + pe - p} \tag{9}$$

The dimensionless transient terminus lake equation, derived from the water balance equation, is now

$$1 - x = \frac{dy}{dx} x \frac{\delta x}{\delta \tau} \tag{10}$$

where the dimensionless area is a function of the dimensionless depth, x = f(y). For a lake with uniformly sloping shores, $x = y^2$ and $dy/dx = 0.5\ x^{-0.5}$, which gives the transient lake equation

$$1 - x = 0.5\sqrt{x}\frac{\delta x}{\delta \tau} \tag{11}$$

An implicit solution of Eq.(11) giving dimensionless time as function of dimensionless area can be found as

$$\tau(x) = \frac{1}{2}\ln\left[\frac{1+\sqrt{x}}{1-\sqrt{x}}\frac{1-\sqrt{x_0}}{1+\sqrt{x_0}}\right] + \sqrt{x_0} - \sqrt{x} \tag{12}$$

where index 0 refers to initial conditions at time 0.

Theoretically, when the climatic parameter $q/(q + pe - p)$ is constant, a new equilibrium lake condition is reached only asymptotically. However, in real life the whether and the runoff varies from year to year, some years are wetter than an average year and some dryer. When the lake level has increased or decreased to a level near the equilibrium one for the new climatic condition, small variations of evaporation, river inflow and precipitation cause the lake level to fluctuate around an equilibrium level. Here the time when the dimensionless lake area, x, has reached a value less than 10%, i.e. $x = 0.9$ or $x = 1.1$, from the equilibrium value unity, is considered to be the time to reach equilibrium, t_{eq}, or in dimensionless form, τ_{eq}. The time to reach equilibrium as a function of initial conditions, $\tau_{eq} = f(x_0)$, as determined from Eq.(12) is shown in Figure 8. The lake area is doubled, x increases from 0.45 to 0.9, which is for example a lake area increase from 10 to 20 km^2, after a dimensionless time of about 0.7. About the same time is required for the lake area to be reduced by 50%, which could be a lake area reduction from 30 to 20 km^2. To go from $x = 10$ to close to unity, lake area 200 to 20 km^2, takes 4 time units. A dimensionless time of about unity is required for obtaining a 10 times increased equilibrium lake area, for example the lake area is increased from 2 to 20 km^2. Thus, the time to reach a new equilibrium condition corresponds to or is in extreme cases a few times the lake response time scale.

For a lake of depth 50 m situated in a river basin in a semi-arid region, where the precipitation deficit, pe-p, is 450 mm yr^{-1} and the annual runoff is 50 mm, the lake response time scale is 100 year. The time scale is smaller for a shallower lake and when the climate is drier. For example, the time scale is 1 year when the lake depth is 1 m, the mean annual runoff is small and the annual precipitation deficit is 1000 mm. In such a climate there is no runoff at all in some years. A very shallow lake is probably dry for part of the year, if lakes exist at all. It is not relevant to discuss the lake response time scale for so extremely dry conditions. As for the semi-arid lake example above, also in Sweden, where the lakes are shallow and the climate is humid, the lake response time scale is 25–100 years, which means that the time to reach a new equilibrium condition should be of the same order. Therefore, it is not possible from known lake levels to draw any quantitative conclusions about precipitation for periods shorter than 50–100 years.

Changed river basin conditions

The basin runoff depends not only on meteorological conditions but also on the physiography of the basin and on the activities within the basin. The runoff, especially the time distribution of the runoff, is different for a mountainous basin compared to a flat basin, and it is different for a forested basin compared to an agricultural basin with clayey soils. However, for the land use to influence the annual runoff and not only the runoff distribution in time, the actual basin evaporation must change due to the land use. Irrigation is a way of increasing evaporation and thus decreasing annual runoff. Therefore, rainfall-runoff simulations were performed assuming that irrigation took place in the summer. For the present South-Swedish conditions, the annual runoff was computed to decrease by only 30 mm compared to when the land was not irrigated, although the simulated irrigation was assumed to be as extensive as possible without lowering the groundwater in a long-time perspective. This small change of the annual runoff is because the river flow is already very low in the summer as is the groundwater level. The effect of irrigation on lakes with outflow may be that their water levels during summer drop below the outlet thresholds, but the effect on their mean water levels is minor. Since the precipitation on a lake and the lake evaporation are the same regardless of whether irrigation takes place or not, the area of a lake without any outflow at all changes linearly with minor changes of the basin runoff, cf. Eq.(4). The 30 mm runoff reduction, although small, corresponds to a lake surface area reduction of about 10% in the present South-Swedish climate. Thus, the annual mean water level of a closed lake are influenced by water use within the river basin.

Table 3. Simulated mean annual water levels (m above sill) of a 0.3 km^2 large lake situated downstream river basins of different area for different annual precipitation. Stage-discharge relation $Q = 1.1\ h^{2.5}$, (V-shaped outlet to a river rapid).

Precipitation mm	Basin area 50 km^2	Basin area 100 km^2	Basin area 200 km^2
200	0.13	0.22	0.33
300	0.28	0.38	0.50
400	0.37	0.50	0.66
600	0.54	0.71	0.94
780	0.67	0.89	1.17
960	0.76	1.01	1.33

Very drastic changes seem to be required to considerably change the water levels in lakes which overflow. The most drastic change one can think of would be a change of the water divide so that the river basin area is changed. It should be possible that erosion in a limited upstream part of a basin may have the consequence that two river basins join into one. If the basin area is doubled, the annual river inflow to a downstream lake is also doubled, and if the basin is increased four

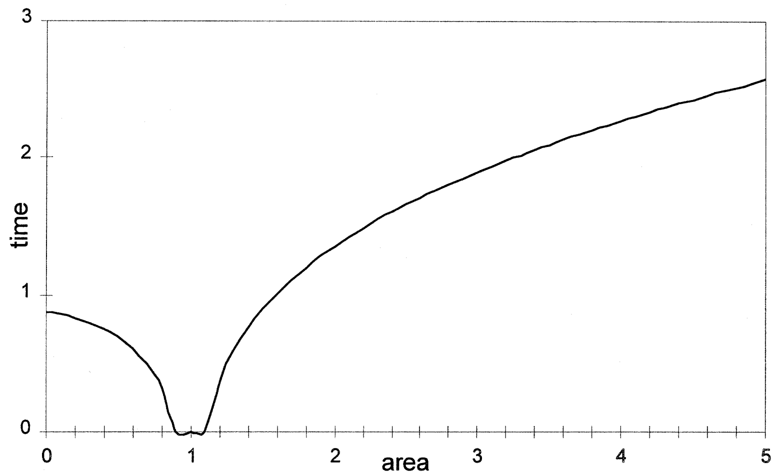

Figure 8. Dimensionless time to obtain equilibrium lake level (defined as 10% deviation from the asymptotic steady-state lake area, which is 5% from the asymptotic lake level) in a lake with uniformly sloping shores as a function of initial lake area relative equilibrium lake area, i.e. Eq.(12).

times, the inflow increases four times. Simulations of lake level variations were done for a lake of 0.3 km^2 area at outlet threshold level when the river basin area was 50, 100 and 200 km^2, respectively. The mean lake levels for different annual precipitation and a potential evaporation of 730 mm per year are shown in Table 3. In the present South-Swedish climate the water level would be 0.4 m higher (0.94 compared to 0.54 m) in a lake downstream the largest basin than in a lake downstream the smallest basin. This difference decreases in a dry climate and increases slightly in a very wet climate. In a closed lake, however, there is a relation, Eq.(4), between lake area and river basin area. In the same climate the lake area is doubled, if the river basin area is doubled.

A basin which is partly glaciated has a different character than a non-glaciated basin. Still, when the glacier neither advance nor retreat, runoff is produced and the runoff does not exceed the precipitation. Thus, the influence of a glacier can not exceed the influence of changing basin area. It was just shown that not even increasing the basin area four times affected the water level in an over-flowing lake very much. Thus, mean annual levels of lakes with outflow are not sensitive even to major changes of conditions in the river basin, whether being due to human impact or natural courses, not even if the water divide of the river basin of the lake is changed. However, human activities within a river basin influence the levels of lakes without outflow.

Estimates of precipitation from reconstructed lake levels

In the previous sections the water balance of overflowing lakes and closed lakes has been discussed in detail. It has been shown that there is a relation between lake levels and annual precipitation but that the relation depends on the seasonal distribution of the rainfall. However, the mean annual water level in lakes with outflow vary in a range of only a meter or less. If there is no river outflow, the lake levels vary much more with varying annual precipitation. Still, the lake level is very much controlled by the river inflow, which in turn depends on the upstream river basin conditions and on the seasonal precipitation distribution. The basin runoff can be determined from rainfall-runoff simulations. By first performing such simulations for different annual rainfalls of given distribution over the year to obtain the river inflow to a lake, and then performing lake routing using the river inflow, the precipitation and the potential evaporation, the mean annual level in the lake can be given as a function of the precipitation.

To illustrate how curves relating lake levels to annual precipitation can look like, two hypothetical lake examples are given. The two lakes, of exactly the same form and dimension, are assumed to constitute 0.3% and 30% of their river basins. The lake shores slope uniformly at 1:50. The outlet is triangular shaped and is a river rapid constituting a hydraulic control section. The basin runoff constituting the inflow to the lakes is the runoff given as function of precipitation in Fig-

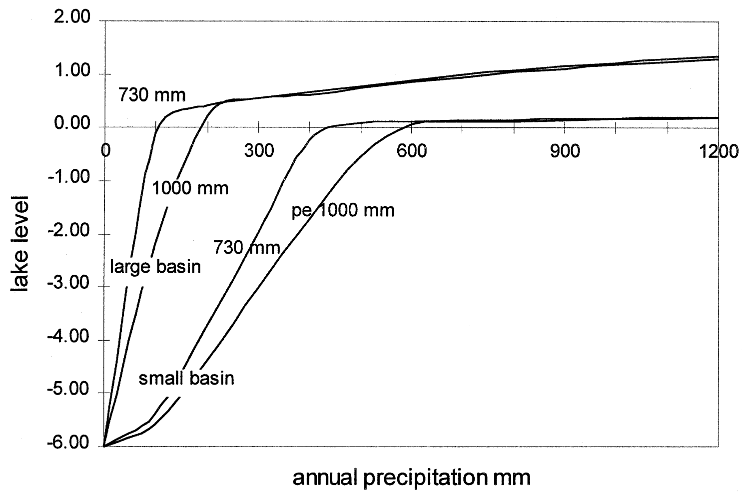

Figure 9. Annual mean lake levels as function of annual precipitation and two different annual potential evaporation values (730 and 1000 mm) for the same 3 km^2 large lake (3 km^2 area at outlet threshold level, triangular shaped river rapid section as outlet, lake shores sloping at 1:50) situated within a small basin of area 10 km^2 and within a large basin of area 1000 km^2, respectively.

ure 7. The computed lake levels for the two lakes as function of annual precipitation are shown in Figure 9.

It is found from the simulations and shown in Figure 9 that the lake in the large river basin should have outflow, when the potential evaporation is 730 mm and the annual precipitation exceeds 100 mm. When the precipitation increases to 1000 mm yr^{-1}, the lake level is about 1 m above the outlet sill. For a lake not to spill over at all when the basin runoff is not very minor, the lake must constitute a large part of its river basin. The lake in the small river basin, which constitutes 30% of the basin, has no outflow for annual precipitation less than 450 mm, when the lake evaporation is 730 mm. When the precipitation is lower than 450–500 mm, the lake level is highly sensitive to the annual precipitation, as is seen from Figure 9. The lake level is reduced to half the depth when the annual precipitation is decreased from 450 to 250 mm.

Simulations were also done for different values of the potential evaporation, which was assumed to correspond to the lake evaporation. As long as the lake remains a closed lake, the lake evaporation influences the lake level. For an annual precipitation of 450 mm, the level of the lake in the small basin was computed to be at sill level when the annual potential evaporation is 730 mm, but 1 m below when it is 1000 mm. For the lake in the large basin changing the annual potential evaporation only affects the computed lake levels, when the annual precipitation is less than 200 mm. The influence of changed potential evaporation relative the influence of changed precipitation is greater the drier the climate is. However, when the upstream basin runoff is not very small, precipitation is generally more important for lake levels than evaporation. This is because evaporation from the vegetation on land can not take place at the potential rate unless much soil water is available, which it is not during dry warm periods.

The time to reach equilibrium conditions in a lake without outflow is directly related to the lake response time scale. For a lake with lake area 3 km^2 at sill level, the response time scale of the lake in the large basin, basin area 1000 km^2, is 30 years, when the annual precipitation is 100 mm and the annual potential evaporation is 730 mm. For the same lake in the smaller basin, basin area 10 km^2, the lake response time scale is 4 years for the same climatological conditions, but about 60 years when the conditions are such that the lake level is near the outlet sill, which corresponds to an annual precipitation of about 500 mm.

The curves of Figure 9 are the result of model simulations. From known lake bathymetry, outlet geometry conditions and river basin area, and using known daily precipitation, temperature and potential evaporation as input, the runoff was computed and then the lake levels and the related lake outflows. From paleolimnological studies it is possible to know the air temperature, from which the potential evaporation can be estimated using the Thorntwaite formula, or by comparing with present day relations solar radiation – temperature –

potential evaporation. Knowing the potential evaporation and thus the lake evaporation, basin area, lake configuration and lake level, the precipitation input into conceptual model simulations can be manipulated to match the reconstructed paleolimnological lake level. It is possible to construct curves as the ones in Figure 9 and from them determine which annual precipitation that corresponds to a certain lake level. However, when computing the annual basin runoff required for computing the lake levels, information about the seasonal precipitation distribution must be at hand or some assumption must be made, since the seasonal precipitation distribution strongly affects the basin runoff. It is only for lake level – climate equilibrium conditions that curves as the ones in Figure 9 relating lake level to annual precipitation are valid. The method suggested here for evaluating precipitation from lake levels can only be used when the climate has been stable for a period approximately as long as the lake response time scale.

Conclusions

Lake levels by themselves give rather little information about local and regional hydrological conditions unless they are related to lake area and river basin area. The annual mean water level in a lake with outflow is not much above the outlet threshold and does not change much, even if the conditions within the upstream river basin change much and the annual precipitation increases or decreases very much. Seasonal fluctuations exceed by far mean annual variations. Little conclusions except that the climate has been quite humid can be drawn from observations of water levels in overflowing lakes. For a lake not to have any outflow, it requires that the inflow to the lake is very small, which means that the lake itself must constitute a significant part of the river basin or that the climate is very dry. In a lake with no outflow, the water level is very sensitive to long-term variations of the climate and also to human activities upstream the lake.

From paleolimnological studies it is possible to reconstruct lake levels and to get information about previous air temperature. From the temperature and the solar radiation the potential evaporation can be determined. Being given information on the river basin and the lake bathymetry, lake levels can be computed for different precipitation conditions using conceptual rainfall-runoff models including lake routing. Thus, it can be found which annual precipitation that corresponds to a certain lake level. This technique can be used for all lakes, but the resolution is high only for lakes which do not have any outflow. The approach requires knowledge of the seasonal precipitation distribution. Unless the lake is small, the method can be used only when the climatic conditions are the same for periods of about 100 years.

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