The water balance of Lake Victoria

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Abstract This paper presents new calculations of Lake Victoria's water balance. Evaporation is estimated using both the Penman formula and the energy balance approach, and sensitivity studies are performed to determine the influence of input data on the estimates. Rainfall over the lake is estimated from catchment rainfall using a relationship between the two that was derived using satellite data. The results, using the reference period 1956–1978, indicate that mean annual rainfall over the lake is 1791 mm, compared to mean annual evaporation of 1551 mm. When compared with lake level changes, tributary inflow, and discharge during this period, there is a resultant imbalance of 19 mm. Adding this amount to the calculated evaporation, we are able to reproduce with great accuracy the lake level changes during the period 1956–1978 utilizing precipitation estimates of this study plus measured inflow and discharge. Sensitivity studies show that the discrepancy in the balance of 19 mm is considerably smaller than the error in evaporation calculations that can be introduced by uncertainties in the input data. Of particular concern is cloudiness. The diurnal cycle of cloudiness is quite different over the lake than at shoreline stations and the total cloud cover over the lake is probably lower than at these stations. A change from 50% cloudiness to 30% can increase evaporation by about 30%. Thus, this study underscores the need for adequate cloud data, sufficient to resolve the diurnal cycle, as well as direct estimates of lake rainfall in assessing the lake's water balance.

Le bilan hydrologique du Lac Victoria

Résumé Ce papier présente de nouveaux calculs du bilan hydrologique du Lac Victoria. L'évaporation est estimée par la formule de Penman et la méthode du bilan énergétique. Une étude de sensibilité a été entreprise pour déterminer l'influence des données d'entrée sur les estimations. Les précipitations sur le lac ont été calculées à partir de données satellitaires utilisant une relation établie entre la pluie sur le lac et celle du bassin versant. Les résultats, basés sur une période de référence allant de 1956 à 1978, indiquent que la moyenne des précipitations annuelles sur le lac est égale à 1791 mm, comparée à une moyenne annuelle de l'évaporation égale à 1551 mm. Une comparaison avec les modifications du niveau du lac, les écoulements et débits pendant cette période indique un surplus de 19 mm. Lorsque cette hauteur est ajoutée à l'évaporation calculée, on arrive à reproduire très exactement les fluctuations du lac de 1956 à 1978, en utilisant les précipitations estimées sur le lac ainsi que les débits et les écoulements mesurés. Les études de sensibilité montrent que l'erreur de 19 mm sur le bilan hydrologique est nettement inférieure à celle induite dans les calculs de l'évaporation en raison des incertitudes sur les données d'entrée. La nébulosité reste un problème majeur car son cycle diurne est sensiblement différent sur le lac et sur les stations côtières, la nébulosité totale sur le lac étant probablement inférieure à celle des stations côtières. Le passage de 50 à 30% de la nébulosité peut augmenter l'évaporation de 30%. Cette étude démontre donc que pour établir un bon bilan hydrologique il est nécessaire d'obtenir des données fiables sur la nébulosité, qui soient suffisantes pour évaluer le cycle journalier, ainsi qu'une évaluation directe des précipitations sur le lac.

INTRODUCTION

The East African lakes have exhibited and left traces of dramatic climatic fluctuations over both historical and recent geologic time (Nicholson & Flohn, 1980, Nicholson,

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1997a, in press). In several lakes, the sedimentary record allows high resolution of climatic fluctuations over several millennia (e.g. Stager *et al.*, 1997; Halfman & Johnson, 1988). Periods of aridity during the late Pleistocene and in the early nineteenth century are ubiquitously evident in the region's lakes. Humid intervals were manifested as major transgressions during the early Holocene and as more recent transgressions in the late nineteenth century and the 1960s. These likewise occurred in lakes throughout East Africa.

Thus, the East African lakes provide a key to the understanding of climate variability in the tropics. Their collective records can yield a spatially and temporally detailed picture of the region's long-term environmental history. Quantification of these lakes' records in climatic terms is also feasible, but this requires mathematical models of their water balances.

Lake Victoria has been of particular interest in the context of water balance because it is the source of the Nile and because of its abrupt fluctuations (Fig. 1) and anomalous hydrological behaviour. Its variability plays a regulatory role for the annual Nile flow, hence any changes in the lake's water balance have significant consequences for riparian countries dependent on Nile water. It has been the subject of several major hydrological investigations, commencing in the early 1930s (e.g. Hurst & Phillips; 1933; de Baulny & Baker, 1970; WMO, 1974, 1981).

The later studies, including those of Flohn & Burkhardt (1985) and Kite (1981), have universally underscored the apparent discrepancies between water balance estimates and lake level fluctuations. In particular, it has been difficult to reproduce the dramatic two-metre rise of Victoria between 1961 and 1964. The problem appears to be in the estimation of rainfall over the lake surface (Kite, 1981; Piper *et al.*, 1986). This is not only the largest term in the water balance, but it is also very difficult to assess. However, it is impossible to completely rule out the inaccuracy in evaporation estimates as the root of the "imbalance". One source of error appears to be associated with the rainfall enhancement over the lake that is produced by the nocturnal lake-induced circulation system (Flohn & Fraedrich, 1966; Fraedrich, 1968).

In this article, we re-examine Lake Victoria's water balance, making two modifications compared to past studies. Most importantly, lake rainfall is estimated from

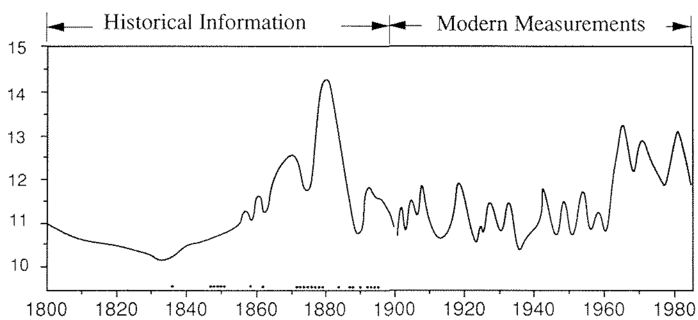

Fig. 1 Fluctuations of Lake Victoria since 1800. Years for which specific references are available are indicated at the bottom with dots. Levels since 1896 are based on modern measurements and earlier years are reconstructed from historical references (from Nicholson, 1997a).

catchment rainfall using a relationship between the two that was derived using satellite data (Ba & Nicholson, in press). Secondly, evaporation is estimated using both the Penman formula and the energy balance approach, and sensitivity studies are performed to determine the influence of input data on the estimates. Our results are compared with past studies and used to reproduce lake-level fluctuations during the period 1956–1978. In a subsequent paper, the model will be utilized to interpret the lake's earlier history in terms of regionally-averaged rainfall.

GEOGRAPHY AND HYDROLOGY OF LAKE VICTORIA

The locations of the major equatorial lakes and rivers are depicted in Fig. 2. Victoria is the largest in Africa, with a surface area of 68 800 km^2. Its mean depth is only 40 m; its maximum depth is about 92 m (Spigel & Coulter, 1996). Lake Victoria receives inflow from 17 tributaries, the largest of which is the Kagera (Howell *et al.*, 1988). However, these contribute less than 20% of the water entering the lake, the rest being provided by rainfall.

Most of the region would be characterized as arid or semiarid; however, around Lake Victoria mean annual rainfall is 1200–1600 mm in most areas (Fig. 3). The

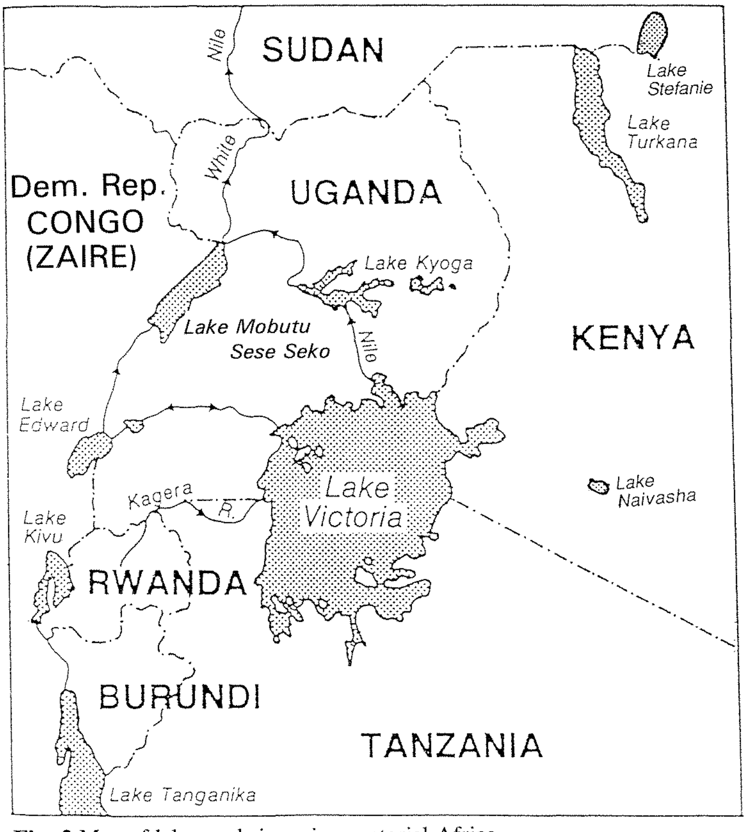

Fig. 2 Map of lakes and rivers in equatorial Africa.

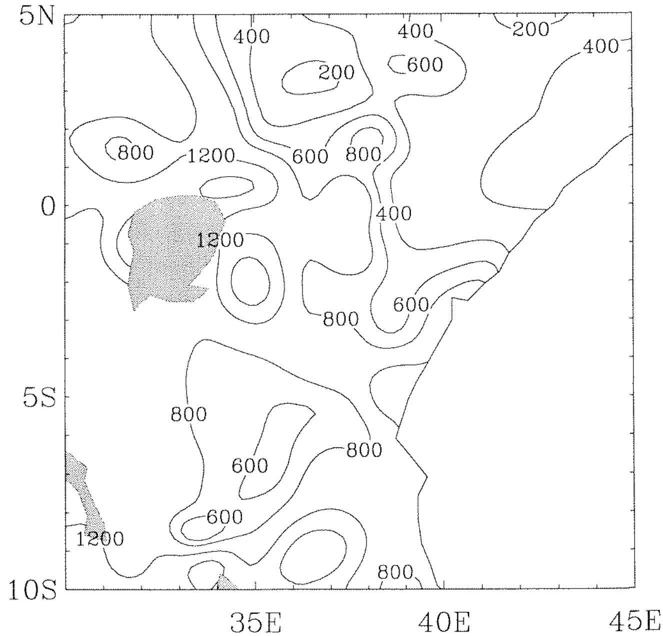

Fig. 3 Mean annual rainfall (mm) over East Africa, based on the period of record (roughly 1930 or earlier to 1994).

most recent water balance estimate, that of Piper *et al.* (1986) gives mean annual rainfall over the lake as 1850 mm and mean annual evaporation over the lake as 1595 mm. Rainfall over the lake is not only the biggest term in its water balance; it is also the most variable. During the period 1970–1974 the range for each term, expressed in multiples of its annual mean, is 0.84–1.22 for rainfall over the lake, 0.71–1.06 for tributary inflow, 0.99–1.02 for evaporation and 0.9–1.14 for outflow (Kite, 1981).

The only outflow is via the White Nile, which exits the lake near Jinja, Uganda. It flows through lakes Kyoga then Albert (now Mobutu Sese Seko) before crossing into the Sudan. Near Khartoum it links with the Blue Nile, which drains the Ethiopian highlands, to form the main Nile flow through Egypt. Thus, Lake Victoria is the source of the White Nile and it also provides 14% of the total Nile flow. The importance of this contribution lies in its relative constancy, a result of the regulating influence of Victoria and the upstream lakes. The rise and fall of these lakes, particularly Victoria, therefore regulates Nile flow. High lake levels caused floods in Cairo in 1964.

Lake Victoria's catchment area is shown in Fig. 4. It totals some 190 000 km^2, and is, hence, nearly three times as large as the lake itself. The catchment includes areas in Rwanda, Burundi, Uganda, Kenya and Tanzania. The northeastern sector (50 000 km^2) is relatively steep and forested, a similarly sized southeastern sector drains somewhat drier and flatter land. In the southwest, the Kagera drains some 60 000 km^2 in the mountains of Rwanda and Burundi, although some of the runoff is lost as it passes through a system of lakes and swamps. The remaining 30 000 km^2 area of the catchment in the northwest contributes little inflow to the lake (Howell *et al.*, 1988).

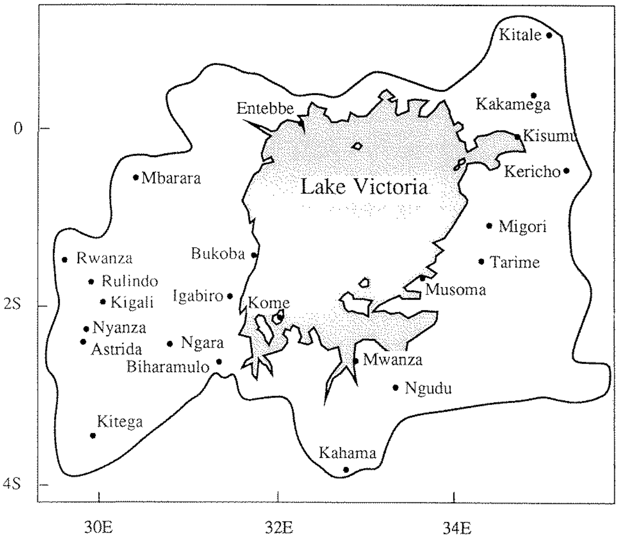

Fig. 4 Approximate catchment area of Lake Victoria and catchment rainfall stations utilized in this study.

PREVIOUS STUDIES OF LAKE VICTORIA

Atmospheric circulation systems

Aerological observations over Lake Victoria stem back to 1908, with the Berson and Elias expedition (Flohn, 1983). Working from a lake steamer buoyed in a harbour, they sent up a balloon with a recording unit to measure winds. When the main balloon broke, the unit fell back over the lake. This expedition provided new insights into the upper air circulation in the equatorial tropics.

Flohn & Fraedrich (1966) further studied the wind regime over Lake Victoria. They described a diurnal circulation system and demonstrated that rainfall over the lake surface is essentially controlled by the convergence associated with the nocturnal land-breeze component of this system. The land breeze regularly produces a giant cumulonimbus cluster lasting until 10:00 or 11:00 h. The cluster is centred during its mature phase over the central and western part of the lake, due to the prevailing easterly winds at the 300–700 mb layer. Consequently, on Victoria's north and west shores (i.e. the Entebbe peninsula and around Bukoba), most rain occurs at night, generally in association with strong thunderstorms. These are observed on somewhat over 50% of all days in the northwest and over the Sese Islands. During the afternoon, the daytime lake breeze diverges and clouds disappear from the lake surface.

The fact that the lake breeze regularly produces cumulonimbus clusters and thunderstorms is related to the thermal instability of the atmospheric boundary layer. The lake, which acts as a storage reservoir of radiational heat, has an average surface temperature (25.4°C) that is about 3.5°C higher than the average air temperature at surrounding coastal stations. This steep lapse rate destabilizes the lower boundary layer. This circulation system is extremely sensitive to changes in the temperature difference between land and water (Flohn & Burkhardt, 1985).

Fraedrich (1971, 1972), using pilot balloon data, constructed a circular convective model, which yielded an areally-averaged rainfall rate of 0.81 mm h^{-1}, rising in the centre of Lake Victoria at the peak hour to 14 mm h^{-1}. Using mean climatological data to establish a frequency of 175 days $year^{-1}$ and a daily duration of 6 h, Flohn (1983) calculated that this circulation produces an average rainfall over the lake of 850 mm. An additional 800 mm would be expected annually through large-scale synoptic processes.

Water balance

The earliest water balance calculations were those of Hurst & Phillips (1933), Hurst (1952) and Merelieu (1961) (see Datta, 1981). Moerth (1967) also studied the relationship between lake level and rainfall. Some of the early estimates pointed out what appeared to be a negative balance, the source of error in the calculations being unknown. A comprehensive measurement and analysis program was started by the WMO survey in 1967. This included establishing gauging stations on all of Victoria's major tributaries to supplement the stations on the Kagera from 1940 and those on four Kenyan tributaries from 1956. Rainfall stations were established around the lake and on islands; index basins were selected to study catchment hydrology and mathematical models were developed to study tributary inflows and lake water balance. This led to the water balance work of de Baulny & Baker (1970) and WMO (1974).

Despite the improved flow of information, the best calculations still resulted in an imbalance between input and outflow. Correcting this problem was one of the tasks of a cooperative project between eight Nile Basin countries and WMO/UNDP (WMO, 1974). That project proposed a standardized scheme for estimating lake rainfall from shoreline and island stations.

This scheme was utilized by de Baulny & Baker (1970) to compile monthly lake isohyetal maps. The method relies upon eight stations around the lake to derive a monthly rainfall series for the period 1925-1969, using monthly weighting coefficients. They assumed a rainfall minimum over the lake (Datta, 1981). Their calculations suggested a mean annual rainfall of 1640 mm. These authors also presented annual inflow series for 1959-1967 for the seventeen tributaries (UNESCO, 1984), but as most of the rivers were not gauged until 1969, these must have been inferred from rainfall. Evaporation was estimated from water balance.

Kite (1981) pointed out that neither the lake's water balance for the 1950s, nor the lake's abrupt rise in the early 1960s could be adequately simulated using a simple water balance model or a lake routing model. The discrepancy was particularly large for the 1961-1964 period, when the net calculated balance of 151 mm was far smaller than the observed 505 mm increase in the height of the lake over this four-year period.

Kite considered the main problem in the calculations to be the lake-rainfall term, but he suspected that previous evaporation calculations were also inadequate. His calculations utilized the series produced by de Baulny & Baker (1970), based on a weighted average of shoreline stations supplemented by data from island stations when available. This method had been utilized since 1970 by the cooperative WMO/UNDP hydrometeorological project (WMO, 1974) and accuracy was given as ±10% (WMO, 1974). When Kite increased the figures of de Baulny & Baker by 25–30%, he could accurately reconstruct Lake Victoria's fluctuations prior to the 1970s using a routing model. Increasing the lake rainfall by 25% in 1977 and 1978 and by 30% in 1979, he could also exactly reproduce the lake's rise during these years. This increase is consistent with the higher values of lake rainfall suggested by Fraedrich (1971, 1972) and by Flohn & Burkhardt (1985).

Piper *et al.* (1986) modified previous estimates of Lake Victoria's water balance in several ways. They extended time series of tributary flow back to 1956, using gauged tributaries to estimate inflow along the ungauged perimeters of the lake, then extended the series further back in time using a rainfall/soil moisture model to estimate runoff. They also used Penman to estimate monthly averages of evaporation, but scaled them on the basis of the annual figure agreed upon by WMO (1974). They computed historic monthly outflows for 1900–1978 using a model of lake levels and outflow. For rainfall they utilized the same eight stations as de Baulny & Baker (1970), but with equal weighting and by averaging normalized departures instead of rainfall totals. The procedure gave equal weight to the variability at each station, irrespective of absolute annual averages. Piper *et al.* (1986) got excellent agreement between predicted and observed lake levels for the 1956–1978 period, and somewhat poorer agreement for a longer time scale (back to 1925) because the tributary inflow model rather than tributary flow was used.

Flohn & Burkhardt (1985) calculated a water balance that took into account the lake effect on rainfall. An increase of rainfall over Lake Victoria had been documented by Flohn & Fraedrich (1966), using several island stations. On those islands, near the northwestern shore, rainfall averaged 2000–2250 mm year^{-1}. Since Fraedrich's model showed that the ascent over the lake has a maximum in its centre, the rainfall from the island stations represents a minimum value that must be attained over the centre of the lake. Data from tiny Nabuyonge Island, in the centre of the lake, indeed shows an average of over 3000 mm year^{-1}. This value is some 30% higher than at any other coastal or island station. A new rainfall map was constructed by combining the Island data with the results of Fraedrich's model (Flohn, 1983). This yielded a revised area-average over the lake of 1690 mm year^{-1}. If this value and a revised evaporation figure of 1470 mm year^{-1} are utilized, a more balanced water budget results. A recent revision of the map (Flohn & Burkhardt, 1985) gave an areal average of 1660 mm, or 1630 mm if the polygon method is used for averaging. The estimated error is 50 mm. This model was able to fully explain the rise of Lake Victoria in the 1960s.

The enhancement of rainfall over the lake and many other aspects of the rainfall regime suggested by Fraedrich's model are supported by the studies of Datta (1981) and Ba & Nicholson (in press). Daily rainfall data show a nocturnal peak in rainfall over the lake's centre and along its western shore, as well as an afternoon peak along

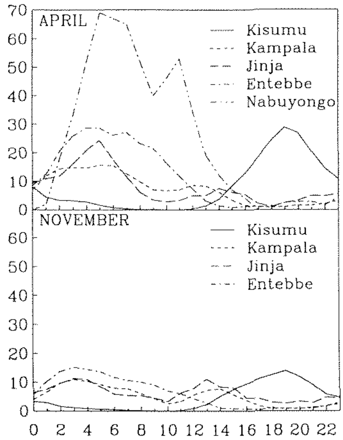

Fig. 5 Hourly amount of rainfall at selected stations during April and November (Ba & Nicholson, in press). The extreme values at Nabuyongo Island are indicative of the enhancement of rainfall by the lake.

the eastern shore (Fig. 5). The 25–30% enhancement is also supported by a satellite analysis of rainfall over Lake Victoria (Ba & Nicholson, in press).

Table 1 presents a summary of the water balance calculations by various studies. The range in values of mean annual rainfall, 1145–1850 mm, partially reflects different periods of calculation, but the estimates are nevertheless widely divergent. Similarly, inflow and outflow vary by a factor of nearly two. The various estimates of evaporation range from 1130 to 1595 mm year^{-1}. The differences largely reflect methods of calculation and assumptions concerning meteorological variables and lake temperature. In general, it appears that the calculations based on energy balance are lower than those produced using the Penman formula. This was confirmed by our own preliminary calculations using both methods.

Table 1 Published estimates of the mean annual water balance (mm) of Lake Victoria, including period of reference, when available.

Reference source	Overlake rainfall	Evaporation	Tributary flow	Lake discharge	Reference period
Hurst (1952)	1420	1350	230	305	
Merelieu (1961)	1145	1130	215–260	305	
de Baulny & Baker (1970)	1630	1523	260	306	1925–1959
WMO (1974, 1981)	1582–1690	1423–1496	238	426	
Hastenrath & Kutzbach (1983)	1650	1500	250	400	
Spigel & Colter (1996)	1450	1370	260	340	
Howell *et al.* (1988)	1810	1593	343	524	1956–1978
Flohn (1983)	1690	1470	280	450	1945–1984
Flohn & Burkhardt (1985)	1630–1660	1470		500	1950–1979
Kite (1982)	1660	1590	420	570	1970–1974
Piper *et al.* (1986)	~1850*	1595	343	~500*	1956–1978
Balek (1997)	1476	1401	241	316	

*estimated from the given curves.

HYDROLOGICAL MODEL OF LAKE VICTORIA

In this section, the basic model is presented and input data for the model are described. Time series of catchment rainfall and evaporation over the lake are derived. Sensitivity studies are then performed with the evaporation model and described in a later section. The mean water balance is calculated for the period 1956–1978, because tributary inflow and discharge are available for that period and because other balance models have been based on this time period.

Basic model equation

The most basic form of our hydrological model of Lake Victoria is a simple water balance between input and output, such that:

$$\Delta H = P_w + \text{INFLOW} - \left(E_w + \text{OUTFLOW}\right) \qquad (1)$$

where ΔH is a change in lake level from the preceding year, input is precipitation over the lake (P_w) plus inflow from the 17 tributaries and output is evaporation over the lake (E_w) plus outflow via discharge to the White Nile. Our initial water balance estimates utilize this equation, calculating the input and output terms as indicated in the following sections. In later work, this equation will be applied to historical fluctuations of the lake, prior to the availability of tributary flow measurements by deriving a regression between land precipitation and inflow.

Rainfall over lake and catchment

The catchment area of Lake Victoria is shown in Fig. 4. There are numerous rainfall stations in the catchment area, and many have records going back to the beginning of the century. However, the stations are concentrated in Kenya and Tanzania. The few stations representing the southwest are generally in Rwanda and Burundi. Consequently, recent data are scarce. Similarly the records from Uganda, covering most of the northwestern catchment, were interrupted during long periods of the 1970s and 1980s.

Our analysis utilizes nineteen rainfall stations in the catchment (Fig. 4) that are available in Nicholson's (1993) African rainfall archive. Stations in that archive were selected on the basis of the length and quality of record and the need to maximize spatial coverage. These stations are considered to be quite representative of a large area because of the strongly coherent patterns of variability throughout this region (Nicholson, 1986). The first year of record and mean annual rainfall at each station is given in Table 2.

The polygon method of spatial averaging of rainfall at individual stations is utilized to calculate catchment rainfall. A comparison with a straight arithmetic average, as used by Piper *et al.* (1986), shows that the technique utilized has little

Table 2 Rainfall stations utilized in this study, first year of data and mean annual rainfall during the period of record.

Station	*First year of data*	*Mean annual rainfall (mm)*
Kakamega	1901	1901
Kericho	1904	1845
Kisumu	1903	1112
Kitale	1921	1180
Migori	1945	1377
Astrida	1928	1155
Kigali	1927	1007
Rulindo	1929	1189
Biharamulo	1922	971
Bukoba	1907	2038
Igabiro	1931	1188
Musoma	1922	825
Mwanza	1902	1027
Ngara	1931	1015
Ngudu	1928	842
Tarime	1933	1465
Entebbe	1901	1531
Mbarara	1912	926
Nyanza	1926	1113

Table 3 Monthly mean rainfall (mm) over Lake Victoria's catchment.

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
90	109	153	210	153	63	47	70	87	111	150	111

effect on the resulting average. For stations with incomplete records, missing data are replaced using a stepwise regression based on all of those stations in the catchment with complete records. The only large gaps in the record are for Astrida (1960–1977), Kigali (1960–1977), and Migori (1931–1944). Mean annual rainfall averaged over the catchment is 1353 mm for the period 1956–1978, with monthly means ranging from 47 mm in July to 210 mm in April (Table 3).

Rainfall over the lake is calculated from the average rainfall in the catchment, weighted by a factor derived from a satellite analysis of rainfall over the lake and over the land (Ba & Nicholson, in press). That analysis (Fig. 6) shows that the frequency of cold clouds is about 25–30% greater over the lake. Estimates of rainfall from the cold-cloud data indicate that the increase over the lake is a relatively constant value of 321 mm. However, this method systematically underestimates the magnitude of rainfall; catchment rainfall calculated from station data for the years 1983–1988 is 32% greater than the satellite estimates. Thus, the difference of 321 mm was increased by 32%, to 424 mm, and this value was added to catchment rainfall to yield an estimate of average rainfall over Lake Victoria.

Figure 7 presents our so-calculated time series of rainfall over the lake. This is compared with the time series of lake rainfall for the years 1956–1978 given by Howell *et al.* (1988) and Kite (1981). Howell does not indicate the source of these

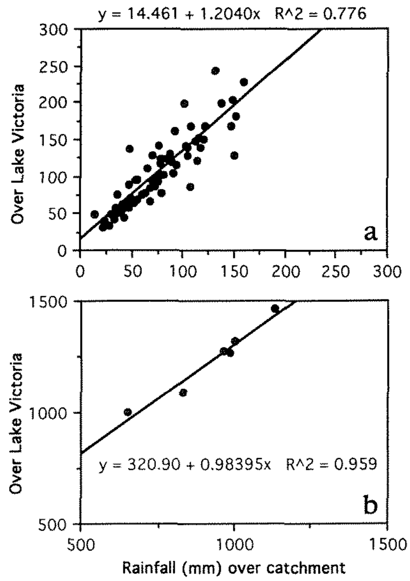

Fig. 6 Scatter plot of rainfall as derived from satellites over Lake Victoria *vs* estimated rainfall over its catchment: (a) spatial averages of the monthly mean computed at the station network for January-December; and (b) spatial averages of annual data computed at the same stations for the years 1983–1990 (from Ba & Nicholson, in press).

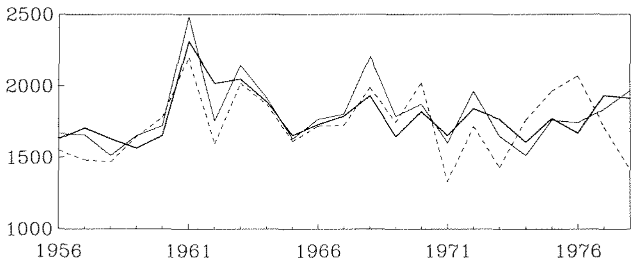

Fig. 7 Time series of rainfall (mm) over the Lake Victoria catchment for the period 1956–1978. (Thick solid line: our study; thin solid line: from Howell *et al.*, 1988; and dashed line: from Kite, 1981).

data, but they are quite similar to those published by Piper *et al.* (1981) in graphical form and based on eight shoreline stations: Jinja, Entebbe, Kalangala, Bukoba, Kagondo, Mwanza, Musoma and Kisumu. Kite's precipitation series is based on the same eight stations plus island rainfall when available. Only five of these stations are in the 19-station network utilized in this study to calculate catchment rainfall. Despite the near independence of our station networks, our results are in remarkable agreement with the data published by Howell *et al.* (1988). The correlation between the two series is 0.87.

Kite's (1981) series shows significant differences from both of the other two series, especially during the 1970s. These differences must be attributed to the data from the island stations that distinguish his series and those of Howell *et al.* (1988).

It may be that lake effect rainfall (and hence island rainfall) does not always vary in phase with catchment rainfall or that the introduction of these data, which are not consistently available, produced inhomogeneities in the resultant time series. If the former is true, this signals an inherent weakness in our method of calculating overlake rainfall from catchment rainfall, since a constant factor is added. However, the strong linear relationship between convective activity over the lake and over the land, as shown by the satellite data (Fig. 6), suggests that rainfall in the two regions does vary in phase, despite the differences in the diurnal distribution in these areas.

Tributary flow

Of the seventeen tributaries providing inflow to Lake Victoria, only four have been gauged for long periods of time. The Kagera River has been gauged since 1940 and the four Kenyan tributaries have been gauged since 1956. The remaining tributaries have been gauged only since 1969.

Table 4 gives the tributary inflow for the period 1956–1978. These were computed by Howell *et al.* (1988) by determining from the later years with complete records the proportion of runoff contributed by the five tributaries operative since 1956 and assuming that proportion remains invariant. This assumption, however, is

Table 4 Mean annual rainfall over Lake Victoria, tributary inflow, discharge, and lake level at Jinja, Uganda, for the period 1956–1978 (from Howell *et al.*, 1988).

Year	Overlake rainfall (mm)	Tributary flow (mm)	Lake discharge (mm)	End of year lake level (m)
1956	1630	288	291	10.91
1957	1707	270	300	11.02
1958	1630	218	294	10.94
1959	1566	199	275	10.84
1960	1654	262	305	10.87
1961	2313	326	307	11.94
1962	2019	539	577	12.39
1963	2049	517	669	12.91
1964	1890	483	753	12.88
1965	1655	260	699	12.48
1966	1731	320	641	12.32
1967	1789	320	564	12.31
1968	1931	487	646	12.58
1969	1645	315	677	12.36
1970	1821	412	661	12.45
1971	1653	301	588	12.17
1972	1839	298	562	12.35
1973	1769	298	573	12.05
1974	1608	313	523	11.97
1975	1773	283	499	12.04
1976	1670	215	521	11.82
1977	1933	326	551	12.13
1978	1914	531	587	12.56

not strictly valid, because the drier catchments to the south provide more variable runoff. The calculated annual inflow varies from 199 mm in 1959 to 539 mm in 1962. The mean for the 23 years is 343 mm, with a standard deviation of 106 mm. These figures have been utilized in our model for the time period 1956–1978.

Outflow

The discharge at Jinja, where the White Nile commences, has been recorded since 1900 (Kite, 1982; Conway & Hulme, 1993). It appears to be directly controlled by lake level (Piper *et al.*, 1986), and hence, can also be predicted from lake level. The discharge between 1956 and 1978 is given in Table 4. This is probably the most accurately known component of Lake Victoria's water balance.

The discharge was originally regulated by Ripon Falls. Since about 1956, it has been regulated by the Owen Falls Dam, constructed in 1954. The flow of the dam, however, has been maintained in such a way as to retain the original lake level/discharge relationship. That relationship is described by a regression, such that:

$$D_i = 0.072H_i + 0.15945H_{i-1} - 2246 \tag{2}$$

where D_i is discharge in year i in mm, H_i and H_{i-1} are the lake levels at the end of the current and previous year, respectively.

Evaporation over the lake

The only direct method of calculating evaporation is eddy correlation. However, due to the technical difficulties this approach presents, the most common methods of estimating evaporation over a lake surface are energy balance and Penman's (1948) combined energy-budget mass-transfer approach. Both rely upon calculations of the radiation balance and heat transfer terms over the lake, which requires a multitude of input data (Table 5) including cloudiness and wind speed over the lake, surface water temperature and surface vapour pressure of the air. Previous water balance calculations make rather simplistic assumptions concerning these and other variables and generally estimate parameters over the lake from land station data, just as in the case of rainfall. Calculations are quite sensitive to these input data and hence, the published values (Table 1) vary by several hundred millimetres per year.

In this study, we assess evaporation utilizing both the energy balance and Penman methods. We then evaluate the sensitivity of the calculations to changes in the most questionable terms, using both methods. However, the evaporation term in our water balance is estimated using the energy balance method. This is in better agreement with other published values of evaporation from Lake Victoria.

Input data Water temperatures of Lake Victoria are given by several authors. Bugenyi & Magumba (1996) indicate that mean monthly maximum water column temperatures for the years 1960 and 1961 were generally on the order of

Table 5 Input data used in the calculations of evaporation from the lake surface.

Month of year	I_0 (W m^{-2})	t_w (°C)	t_a (°C)	e_a (mb)	u_2 (m s^{-1})	LW (W m^{-2})	R_{net} (W m^{-2})
January	418.2	25.5	22.6	19.7	1.90	63	139
February	431.3	26.3	22.8	19.9	1.95	63	145
March	433.2	26.3	22.9	20.5	1.97	62	147
April	418.7	25.8	22.5	21.3	1.87	60	142
May	393.0	25.4	22.2	21.0	1.83	60	129
June	377.5	25.3	21.8	19.4	2.08	63	119
July	380.9	24.7	21.3	18.3	2.08	65	118
August	403.7	24.8	21.6	18.7	2.00	65	130
September	424.5	24.7	22.1	19.0	2.05	64	141
October	429.8	24.9	22.7	19.5	1.98	63	144
November	419.7	25.4	22.7	19.7	1.90	63	139
December	411.4	25.7	22.4	19.9	1.78	63	136

Notes:

I_0: solar radiation received on a horizontal plane at the upper edge of the atmosphere (Black *et al.*, 1954). Interpolated to 1°S for Lake Victoria;

t_w: water temperature derived from Talling (1969);

LW is the long wave radiation of the lake;

R_{net} is the net radiation received by the lake surface;

The remaining terms t_a, (air temperature), e_a,(air vapour pressure), u_2 (wind speed at 2 m high over the lake surface) are the average values of the six lakeshore stations.

25.5–26.6°C, with mean minima on the order of 24–25°C in most months. The diurnal range is considerably lower during the months of February–June than at other times. Temperatures were 0.5–1°C higher during the period 1990–1992, with most of the change apparent in maximum water column temperatures. Ochumba (1996) indicates measured monthly means as high as 27 or 28°C in the early months of the year at numerous locations on the lake. Talling (1969) shows surface temperatures ranging from about 24.5–26°C during the course of the year, and temperatures of the order of 24°C at 55–60 m depth. A similar temperature regime is indicated by Beadle (1974). Talling's surface temperatures are used in this study, as they probably best represent the temperatures during the period of model validation, 1956–1978.

We have averaged climatological data from six surface stations along the lakeshore (Entebbe, Jinja, Kisumu, Musoma, Mwanza and Bukoba) for estimates of wind, vapour pressure and air temperature (Table 5). Mean monthly temperatures at individual stations range from 21–24°C, while the mean of the six stations ranges from 21.3°C in July to 22.9°C in March. Mean vapour pressure at these stations ranges from 18.3 mb in July to 21.3 mb in April. Winds are relatively low at all stations, generally on the order of 1–3 m s^{-1} at a height of 2 m. Higher speeds might be anticipated over the lake itself because of lower friction. On the other hand, the daytime subsidence over the lake and the divergent surface winds associated with it might reduce the wind speeds, relative to winds over the land. Nevertheless, the average wind speed from the shoreline stations in Table 4, 1.95 m s^{-1}, is in close agreement with a mean of approximately 2 m s^{-1} derived by Ochumba (1996) from ten stations over the open lake and in sheltered bays of the lake.

The most difficult parameter to estimate is cloudiness over the lake and, unfortunately, both methods of calculating evaporation are extremely sensitive to this parameter. Cloudiness is available for the stations in Table 5, but the estimates, which are often on the order of 5 or 6 octals, appear to be too high to represent cloudiness over the lake. Instead, the mean fractional cloud cover over the lake has been set to 0.5, a value taken from the Atkinson & Sadler (1970) atlas and used in the water balance study of Hastenrath & Kutzbach (1983).

Admittedly, this may still be too high. The shortwave radiation balance is driven by daytime cloudiness, which is extremely low over the lake, due to the subsidence associated with the lake-land breeze system (NOAA-USAF, 1971; Kayiranga, 1991; Ba & Nicholson, in press). Because of the diurnal cycle of clouds, intense sunshine is generally restricted to between the hours of 11:00 and 16:30 (Bugenyi & Magumba, 1996). For the longwave balance, mean diurnal cloudiness is of relevance and this may be close to 0.5, being possibly as high as 100% during much of the night and near zero during much of the day. Both mean cloudiness and its diurnal cycle have a strong seasonal dependence as well.

Radiation balance calculations Radiation balance is calculated from empirical formulae for both shortwave and longwave radiation. Top-of-the atmosphere solar insolation at the latitude of the lake is approximated from a table in Black *et al.* (1954). Cloudiness is taken into account using the formula:

$$G = I_0 \times \left(0.803 - 0.34c - 0.458c^2\right) \tag{3}$$

from Black *et al.* (as quoted in Dunne & Leopold, 1978). Here G is global radiation, I_0 is insolation and c is fractional cloud cover. This equation yields values ranging from 182 W m^{-2} in June to 209 W m^{-2} in March, with an annual average of 199 W m^{-2}. Water surface albedo is evaluated from Anderson's equation (WMO, 1966), which gives slight month-to-month variations with changing solar zenith angle but values close to 0.07 in all cases. Hence, 0.07 is adopted as the lake's albedo. Net shortwave radiation is then calculated as:

$$SW = G(1 - \alpha) \tag{4}$$

where α is the albedo.

The net longwave radiation is given by Budyko (1974) as:

$$LW = \varepsilon\sigma\left(t_w + 273.15\right)^4\left(0.39 - 0.05\sqrt{e_{sat}}\right)\left(1 - 0.53c^2\right) \tag{5}$$

where ε is the emissivity of water, $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan-Botzmann constant, t_w is the monthly water temperature, e_{sat} is surface vapour pressure and c is fractional cloud cover. Water surface emissivity is set at 0.96 (Sellers, 1965). The saturation vapour pressure of the water surface e_{sat} is calculated as described in the following section for the monthly mean surface temperatures t_a given in Table 5. Calculated values of net longwave radiation range from 60–65 W m^{-2}. The total net radiation received by the water surface is then derived as:

$$R_{net} = SW - LW \tag{6}$$

This would mean that total net radiation is on the order of 110–135 W m^{-2}.

The shortwave and net radiation estimates are in excellent agreement with satellite-based assessments made by Ba & Nicholson (submitted) that take into account atmospheric corrections. Their results indicate that shortwave irradiance at the surface in the vicinity of Lake Victoria is on the order of 200–225 W m^{-2}, with total net radiation on the order of 125–150 W m^{-2} throughout the year. Measured solar radiation at the surface likewise validates our calculations (Shahin, 1985). Annual averages for the stations Jinja, Entebbe, Kisumu, Bukoba and Mwanza range between 185 and 230 W m^{-2}.

The energy balance approach The essence of the energy balance approach is that, under steady-state conditions, the net radiation at the surface is balanced by latent and sensible heat transfer, or:

$$R_{\text{net}} = S + LE \tag{7}$$

where S is sensible heat transfer between the atmosphere and lake and LE is the latent heat used to evaporate water from the lake. This equation is rewritten through the use of the Bowen ratio, defined as:

$$B = \frac{S}{LE} \tag{8}$$

The sensible and latent heat fluxes are commonly evaluated using "bulk formulae" that relate these fluxes to the average differences in temperature and vapour pressure between the water surface and some fixed height above it. Strub & Powell (1987) have demonstrated experimentally the validity of this approach for calculating climatological heat fluxes over oceans and lakes. In this study, the Bowen ratio is estimated by the bulk formula:

$$B = \frac{0.61p(t_w - t_a)}{1000(e_{\text{sat}} - e_a)} \tag{9}$$

where p is the atmospheric pressure at the elevation of the lake in mb; t_w and t_a (°C) are the water surface and air temperature, respectively; e_a is the atmospheric vapour pressure in units of mb; and e_{sat} is the saturation vapour pressure at the temperature t_w:

$$e_{\text{sat}}(t_w) = 6.11\exp\left(\frac{17.3t_w}{t_w + 237.3}\right) \tag{10}$$

Using the data in Table 5, these equations yields a value of 0.15 for the Bowen ratio over Lake Victoria. With the help of the Bowen ratio definition, equation (8) can be rewritten to obtain the lake evaporation, which is then expressed as:

$$E = \frac{R_{\text{net}}}{(1+B)L} \tag{11}$$

The resultant values of evaporation from Lake Victoria, using the input data in

Table 6 Monthly evaporation over Lake Victoria, as calculated by the energy balance and Penman methods.

Month	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
Radiation	135	127	144	132	123	109	113	126	135	145	132	129	1551
Penman	150	145	162	146	138	130	135	144	148	154	145	146	1743

Table 5, are given in Table 6. Monthly values range from 109 mm in June to 145 mm in October. Annual evaporation is 1551 mm, which is in good agreement with most of the previously published calculations indicated in Table 1.

The Penman approach Penman (1948) combined the energy budget approach with the concepts of mass-transfer and a formula for potential evapotranspiration. This approach has been used in assessing the water balance in numerous studies of lakes such as, for example, Lake Bosumtwi in Ghana (Turner *et al.*, 1996) in Penman's formula, as quoted by Dunne & Leopold (1978) and Dingman (1994), estimates evaporation E as:

$$E = \frac{R_d\Delta + \gamma E_a}{\Delta + \gamma} \tag{12}$$

where R_d is the net radiation equivalent to the amount of water that can be evaporated in unit of cm day^{-1} with 1 cm^2 cross-section:

$$R_d = \frac{R_{\text{net}}}{\rho L} \tag{13}$$

and Δ (in mb °C^{-1}) is the slope of the curve relating saturation vapour pressure to temperature:

$$\Delta = \frac{\mathrm{d}e_{\text{sat}}}{\mathrm{d}t} = \frac{25083}{\left(t_w + 237.3\right)^2}\exp\left(\frac{17.3t_w}{t_w + 237.3}\right) \tag{14}$$

The psychrometric constant γ is given by:

$$\gamma = \frac{c_a P}{0.622\lambda_v} \tag{15}$$

where the heat capacity of air c_a = 0.24 cal g^{-1} °C and P is the atmospheric pressure at the elevation of the lake being studied. The latent heat of vaporization in (cal g^{-1}) is approximated by:

$$\lambda_v = 597.3 - 0.564t_w \tag{16}$$

The term E_a describes the contribution of mass transfer to evaporation. The empirical formula is thus:

$$E_a = (0.013 + 0.00016u_2)(e_{\text{sat}} - e_a) \tag{17}$$

where u_2 is the wind speed at 2 m height over the lake expressed in km day^{-1}.

The resultant values of evaporation from Lake Victoria, using the input data in Table 5, are given in Table 6. Monthly values range from 130 mm in June to 162 mm in October. Annual evaporation is 1743 mm. This is nearly 200 mm higher than with the energy balance method and it is higher than any of the previously published calculations indicated in Table 1. Both Penman and energy balance estimates show a double maximum in evaporation, peaking in both March and October and reaching an absolute minimum in June.

Sensitivity studies To determine the influence of wind speed, cloudiness, air and water temperature and humidity on evaporation, we have performed the calculations using a range of values for each term. Wind speed is varied from 0–6 m s^{-1} and cloudiness from 0–100%. Air and water temperatures, as indicated in Table 5, are varied by +4°C to -4°C in increments of 1°C and vapour pressure in the Table is varied by +4 mb to -4 mb. Net radiation is varied from +30% to -30% of the values in Table 5. The results are shown in Fig. 8.

The models show the least sensitivity to vapour pressure, one of the more accurately known quantities. As it changes by +/-4 mb, evaporation calculated by the energy balance method varies from 1585 to 1517 mm, while that from Penman varies from 1726 to 1810 mm. With the former method, evaporation is decreased as humidity increases, but the opposite occurs with Penman.

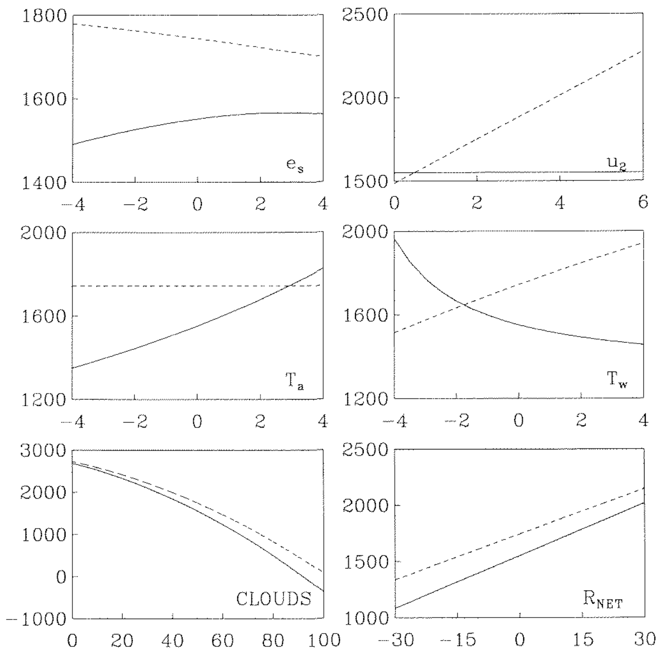

Fig. 8 The results of sensitivity studies: the effect of systematic changes in saturation vapour pressure, wind speed, air and water temperatures, cloudiness and net radiation on evaporation. (Solid lines: energy balance method; dashed lines: Penman method. Vertical axis is evaporation (mm $year^{-1}$). For vapour pressure and temperature, the horizontal axes represent changes with respect to the values in Table 5. For the remaining variables, the axes represent absolute values.)

The energy balance method is insensitive to wind speed, but the Penman method is exceedingly sensitive to it, with evaporation increasing with wind speed. As the mean wind speed at two metres varies from 0 to 5 m s^{-1}, annual evaporation in the Lake Victoria example changes from 1509 to 2297 mm. This range may be somewhat unrealistic, but even within the range of 1–3 m s^{-1} it varies from 1641 to 1903 m s^{-1}.

Both methods are less sensitive to temperature. Penman shows no sensitivity to air temperature, but with the energy balance method evaporation increases markedly with air temperature. As the mean temperatures in Table 5 are progressively altered by -4°C to +4°C, evaporation changes from 1371 to 1855 mm. For water temperature, the functional dependence is the opposite in the two methods. With Penman it changes over this range from 1539 to 1975 mm $year^{-1}$, but with the energy balance method it decreases from 1994 to 1480 mm $year^{-1}$ as water temperature is changed by +4°C to -4°C. For Lake Victoria, however, the uncertainties in air and water temperatures are probably relatively small, at most 1°C. This corresponds to a range of evaporation of 1520–1577 mm $year^{-1}$ for air temperature. For water temperature, the range is 1624–1523 mm $year^{-1}$ with the energy balance method and 1717–1825 mm for Penman. The increase in lake temperature of 0.5–1°C between the 1960s and 1990s, as noted above, would have altered mean annual evaporation by about 30–60 mm.

Both methods are extremely sensitive to cloudiness, a parameter that is not accurately known for Lake Victoria, and to net radiation, a parameter that is strongly dependent on cloudiness. As cloudiness changes from 0–100%, mean annual evaporation changes from around 2600 mm to as low as 800 mm with the energy balance method and 1100 mm with Penman. We have assumed, as did many other studies, that the mean cloudiness, based on shoreline stations, is about 50%. In view of the diurnal cycle in cloudiness associated with the lake and land breeze system and the subsidence over the lake during the day, the mean could realistically be as low as 20–30%. This would increase evaporation by as much as 400–600 mm $year^{-1}$. Hence, the uncertainty in cloudiness creates a large uncertainty in estimates of annual evaporation over the lake. The sensitivity of evaporation to net radiation is similar to that with respect to cloudiness. The range, as net radiation varies from +30% to -30%, is from 2017–1086 mm using the energy balance method and from 2147–1339 mm using the Penman formula. In both cases, evaporation increases linearly with net radiation, somewhat more strongly with the energy balance approach.

RESULTANT WATER BALANCE

The mean annual values of the principal water balance parameters for the years 1956–1978 are given in Table 7. Rainfall over the lake is 1791 mm, evaporation is 1551 mm, discharge is 524 mm and inflow is 338 mm. During this interval, the net lake level change was an increase of 1680 mm, corresponding to an increase in "storage" of 73 mm annually. Thus, the difference between our calculated water input (2129) and output (2148) is 19 mm. It is noteworthy that the lake rainfall

Table 7 Mean annual water balance (mm) of Lake Victoria 1956–1978, as calculated by this study.

Over lake rainfall	Tributary flow	Evaporation from lake	Discharge from Jinja	Lake level change
1791	338	1532	524	73

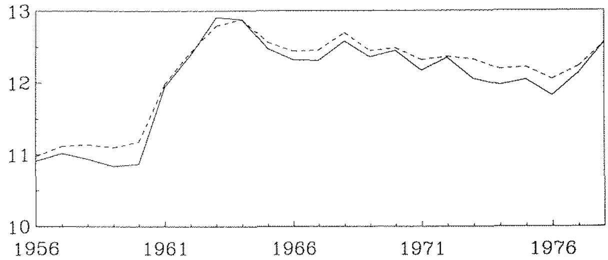

Fig. 9 The level in metres of Lake Victoria 1956–1978 as estimated by our model (solid line) and as measured at Jinja, Uganda (dashed line).

estimates in Howell *et al.* (1988), that exceed ours by 19 mm, would give an exact balance.

In order to avoid cumulative error in the prediction of lake levels from water balance, a complete balance must be achieved for the test period 1956-1978. Since the least accurate term is evaporation, and 19 mm is considerably smaller than the potential errors in evaporation due to uncertainties in input data, the model is balanced by decreasing evaporation from 1551 to 1532 mm year^{-1}.

Figure 9 shows predicted and observed lake levels measured at Jinja, Uganda, during the period 1956–1978. There is excellent agreement and the increase in lake levels in the early 1960s is very precisely reproduced. Greatest discrepancies are in the years just prior to the rise. The correlation between predicted and observed levels is 0.99.

SUMMARY AND CONCLUSIONS

This study differs from previous water balance studies of Lake Victoria in two ways. First, lake rainfall is assessed from a large number of stations distributed over the catchment, in conjunction with satellite estimates of rainfall directly over the lake itself. Second, the sensitivity of the lake evaporation to climatological input data is evaluated.

Our calculations suggest that mean annual rainfall over the lake for the period 1956–1978 was 1791 mm. The resultant annual time series of lake rainfall is remarkably similar to previous estimates based on far fewer stations and without direct estimates of lake rainfall. It therefore validates these previous estimates, generally based on complex statistical models.

This study demonstrates significant differences in evaporation calculated via energy balance considerations and via the Penman formula. There are shortcomings to both approaches. However, a comparison of evaporation calculated via the energy balance method, our rainfall estimates, and measured tributary flow, discharge and lake levels for the period 1956–1978 indicates a very precise balance. The discrepancy of 19 mm is utilized to adjust evaporation estimates to achieve a balance for this period. When this is done, there is excellent agreement between lake level fluctuations predicted from the water balance model and those measured at Jinja.

This discrepancy is considerably smaller than the error in evaporation calculations that can be introduced by uncertainties in the input data. Of particular concern is cloudiness. The diurnal cycle of cloudiness is quite different over the lake than at the stations along the shore and the total cloud cover over the lake is probably lower than along the shore. A change from the assumed 50% cloudiness to 30%, perhaps a more realistic figure, can increase evaporation by about 30%. A more precise estimate of cloudiness requires data directly over the lake, during both day and night.

Earlier studies have tended to manipulate rainfall data to balance Lake Victoria's water budget. Kite (1981), however, suggested that there is considerable uncertainty in the evaporation estimates as well. Our study suggests that this, rather than rainfall, is in fact the most imprecisely known aspect of the water balance. Our study also underscores the need for adequate cloud data, sufficient to resolve the diurnal cycle, and for direct estimates of lake rainfall in resolving the lake's water balance. This is true of the other Great Lakes of Africa as well, such as Tanganyika, Malawi and Chad.

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