Modelling of the hydrology of the Okavango Delta

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Table of contents

Table of contents ........................................................................................................... 1
List of figures .................................................................................................................. 1
List of tables .................................................................................................................... 2
1. Introduction .................................................................................................................. 3
2. Characteristics of the hydrological system of the Okavango Delta ....................... 4
3. Previous modelling ..................................................................................................... 6
4. The hydrological model ............................................................................................. 8
   4.1 Adopted modelling concept .................................................................................... 8
   4.2 Reservoir model .................................................................................................... 8
      Basic reservoir setting .............................................................................................. 9
      Reservoir links .......................................................................................................... 9
      Surface water-groundwater interactions ................................................................ 9
      Evaporation ............................................................................................................. 11
      Rainfall .................................................................................................................... 12
      Calculation procedure ............................................................................................ 12
      Derivation of model parameters ............................................................................ 12
      Calibration of the model ........................................................................................ 13
   4.3. GIS model .......................................................................................................... 14
5. Results and discussion ............................................................................................. 16
   Outflows ...................................................................................................................... 16
   Flood size ................................................................................................................... 16
   Boro flow regimes ...................................................................................................... 16
   Shifts & changes in the system .................................................................................. 17
   Active and passive storage ....................................................................................... 17
   Spatially distributed flood characteristics ................................................................ 18
   Channel flows vs distributary flows .......................................................................... 18
   Water balances .......................................................................................................... 18
6. Summary and conclusions ....................................................................................... 19
Tables ............................................................................................................................. 20
Figures ........................................................................................................................... 22
References ....................................................................................................................... 33

List of figures

Fig. 1 Okavango Delta and its main features .................................................................. 22
Fig. 2 Basic model setting a) units represented by surface water reservoirs, b) schematic for unit
    reservoir setting (SW- surface water reservoir, FGW – floodplain groundwater reservoir, IGW –
    island groundwater reservoir) .................................................................................... 23
Fig. 3 Comparison of $e_{ref}$ for Maun and Nxaraga and $e_{est}$ from an eddy covariance system at papyrus
    and seasonal floodplain site (location of all the sites in Fig. 1) ............................... 24
Fig. 4 Volume-area curves for a small floodplain, used in the previous models and derived from DEM
    for large units ............................................................................................................. 25
Fig. 5 Observed and simulated discharges at Maun and Toteng .................................. 26
Fig. 6 Observed and simulated a) flow duration at Maun and b) flood size exceedance in Boro
    distributary ............................................................................................................... 27
Fig. 7 Observed and simulated flood sizes in Boro, Xudum, Thaogoe and Mboroga distributaries... 28
Fig. 8 Observed water levels and flood extent for the Ngoga unit .................................. 29
Fig. 9 Flood frequency map for the 1990-2000 period (observed, simulated and difference) ..... 30
Fig. 10 Island groundwater storage ($V_o$) and floodplain-island groundwater flow ($Q_{ref}$) for Boro
    distributary as simulated by the model .................................................................. 31
Fig. 11 Comparison of observed channel flows and modelled distributary flows for two sites:
    Mmadinare and Gaenga (location in Fig. 1) .............................................................. 32
List of tables
Table 1 Model performance criteria ................................................................. 20
Table 2 Long term (January 1969-December 2002) water balance of the Okavango Delta .......... 21
1. Introduction

The Okavango Delta is a large wetland dominated by an annual flood from the Okavango River. During such a flood event, the extent of the inundated area increases from about 5000 km\(^2\) to 6000-12000 km\(^2\), depending on the size of the flood. The Delta is located in semi-arid NW Botswana, where highly seasonal rainfall (November-March) is in the order of 300-500 mm·a\(^{-1}\) and potential evaporation (corrected pan A) amounts to 2100 mm·a\(^{-1}\). The more than 600 km distance from the headwaters of the Okavango River and the low topographic gradient of the alluvial fan (1:3500), cause a delay of the annual flood in the system. As a result, the flooding occurs in the distal part of the Delta only during the late dry season (August-October). Such a setting creates an environment with a very specific ecology. Murray-Hudson et al. (submitted), explains that the frequency of annual floods and flood duration are the principal drivers of the ecosystem. Both the feeding river and the Delta itself have a high potential for water resource development (Kgathi et al., submitted; Andersson et al., submitted).

The hydrological model presented here has been created in order to provide a realistic tool for the analysis of impacts from upstream abstractions and developments that modify the hydrograph of the Okavango, as well as possible impacts from climatic change. This analysis is particularly important for the delicate link between the hydrological behaviour and the ecosystem (Murray-Hudson et al., 2005; submitted). In addition, the model enhances understanding of the hydrology of the Okavango Delta and provides insight into the possible effects of long term changes to the system.
2. Characteristics of the hydrological system of the Okavango Delta

The hydrology of the Delta has been frequently described in the literature (e.g. Gumbrecht et al., 2004b; McCarthy et al., 1998; Gieske, 1997) and therefore only the main elements are presented here. The flooding in the Okavango Delta is primarily caused by the annual flood wave arriving from the Angolan part of the Okavango river basin. At Mohembo, the Okavango River enters a 10-15 km wide, 150 km long, flat-bottomed valley called the Panhandle. At the mouth of the Panhandle, the river enters a broad, essentially unbound flat area forming the apex of the alluvial fan or the Delta proper. That area is called the central swamp. There, flow separates into several fan-like oriented distributaries, each of which is formed by a 10-30 km wide system of interconnected floodplains, interspersed with islands of various sizes. In response to the annual flood wave, inundated area increases and decreases in each of the distributaries, primarily along the longitudinal slope of the alluvial fan. The dynamics of the annual flood varies between the systems. The eastern distributaries are characterized by a rather small seasonal and interannual variation and a relatively large area that is permanently inundated, while the central and western distributaries are much more dynamic (see Fig. 7).

Flow in the system takes the form of a combined floodplain/channel flow with channel flow velocities in the order of 0.4-0.8 m s\(^{-1}\), and floodplain velocities being typically less than 0.01 m s\(^{-1}\). Channels in the Okavango Delta are flanked by vegetation (papyrus and reeds) and as a result, water is freely exchanged between channels and surrounding floodplains. Channels are relatively well defined in the Panhandle and in the central swamp, and there they are mostly losing water to the floodplains. In the downstream, seasonally inundated parts, however, channels lose their continuity and integrity, and take the form of a set of often disconnected stretches of varying width and characteristics. Wolski et al. (2005) show that channel flow dynamics changes in a non-consistent fashion along a distributary, indicating a major role of ungauged off-channel flows in flooding. This tallies with conclusions of McCarthy and Ellery (1997) of channels in the system playing the role of distributing sediment rather than water.

The alluvial fan of the Okavango Delta is built primarily of so-called Kalahari Sands, only to a small extent modified by precipitated clayey material. As a result, the substratum is well permeable (typically 1-20 m d\(^{-1}\)), facilitating exchange of surface and groundwater. Infiltration at the arrival of the flood is therefore rapid – in order of 100-400 mm d\(^{-1}\). As a result of the continuous uptake of shallow groundwater by island vegetation, local groundwater flow systems develop between floodplains and islands. These groundwater flow systems are very dynamic and controlled by island transpiration Wolski and Savenije (2004). In a typical small to medium-sized island (100-1000 m in width), groundwater storage, depleted during a no-flood season or even a series of no-flood years, can be
refilled during one flood. In case of larger islands (>1 km width), this process can last several years. At the two small floodplains from which water balances are available (Dinçer et al., 1976; Ramberg et al., 2004), on an annual basis, infiltration exceeds evaporation from the water surface by a factor of 6 to 9.

Local rainfall falls during the annual flood minimum. High permeability of the substratum together with flat regional topography and the specific topography of islands, which are characterized by elevated fringes with depression in the middle (McCarthy and Ellery, 1994), cause the surface runoff from outside of floodplains being virtually non-existent. Rain-induced infiltration-excess runoff or saturation runoff is possible in the dry floodplains, however, appears to be of limited regional significance and forms mostly small isolated pools, rather than cause extensive areal inundation. Nonetheless, local rainfall is a significant factor influencing interannual variability of flood magnitude, but rather in the way of "pre-wetting" the system before the arrival of the Okavango River flood wave, than by inducing flooding itself. Only in extremely high rainfall years (>800 mm·a⁻¹) rain-induced floods can occur.
3. Previous modelling

In the past, several hydrological models of the Delta were created. These were reservoir models and they simulated either a part or the entire system. These were calibrated against observed outflows from one of the distributaries, the Boro. The old models have two major drawbacks that prevent them from being applicable for the purpose of this study.

Firstly, the models did not allow for the determination of ecologically significant variables, such as duration and frequency of floods. The early models (Dinçer et al., 1987; SMEC, 1990; Scudder et al., 1993) were merely meant to provide information on the amount of water that can technically be abstracted from the Delta. Environmental effects of technical interventions into the system were not explicitly addressed, and the only ecologically sound variable used was the area of inundation. The WTC (1997) model addressed the issue of effects of upstream abstractions on the flooding, but that was done in a simple manner: average reduction in outflow and average reduction in flooded area were the indices used. Similar indices were used by Gumbrecht et al. (2004b) in their regression model for the maximum annual flood.

Secondly, the models were not conceptually convincing. During calibration of these models, a major problem was encountered: during a part of the observation period (1974-1981), the models consistently underestimated outflows (by 20-50%) of the Boro distributary, while performing well during the remaining part. Dinçer et al. (1987) and SMEC (1990) explained this as an unsystematic change in the flow regime of the Boro distributary, and imposed an empirical change in model parameters to represent it. That change caused the model to simulate additional inflow to the Boro at the expense of the neighbouring distributary, the Xudum. The authors argued that the change in hydraulic properties affecting the distribution of flow in the system could have been caused either by tectonic activity, by sedimentation or by vegetation change. Since all three processes are recognized as possibly affecting the distribution of flow in the Okavango Delta (McCarthy et al., 1988), such explanation had some degree of likelihood. The Scudder et al. (1993) and WTC (1997) models considered the discrepancy being a result of errors in rainfall data. Based on the analyses of double mass curves, 14 different correction factors, as large as 1.2, were applied to the 70 years of record of the two rainfall stations, Maun and Shakawe, data from which were used as input to the model. That improved overall performance of the model, but the outflows were still underestimated during the problematic years. Gieske (1997) noticed that the period when the outflows were underestimated coincides with a period of high values of a long-term antecedent rainfall index (cumulative rainfall departure - CRD) that is used in estimation of groundwater recharge. In his model, Gieske related one of the model parameters that described transfer of water between reservoirs to that index, and was able to simulate the whole time series of outflows rather well. He therefore concluded that the increased outflows could have been caused by a groundwater recharge related process, although he
did not identify it. His solution stipulated that the increased outflows in the Boro distributary occurred at the expense of the neighbouring Xudum distributary.

Recently, a distributed model based on the MODFLOW code was developed (Bauer, 2004). In spite of its distributed nature, the model uses spatially-uniform parameterization of flow in both the surface water and the groundwater layers. As such, it accounts only for the effects of the geometry of the flow system flood propagation. The model was calibrated against flood maps (McCarthy et al., 2004), and the 30-year long time series of outflows from the system was not simulated at all. In spite of that, that model is a significant step towards building a physically sound mathematical representation of the Okavango Delta system. However, more work has to be done on independent parametrization in order to make distributed models more realistic.
4. The hydrological model

4.1 Adopted modelling concept

Since remote sensing derived flood maps are available (McCarthy et al., 2004), we developed a tool that integrates reservoir modelling and GIS-modelling. The reservoir model provides information on the flood extent in rather large units and the fluxes between them. Subsequently, the GIS model allows for the translation of the unit flood size into a flood distribution map, and hence spatially distributed flood characteristics, which can be used to determine eco-system responses.

Gieske's (1997) explanation for the apparent non-homogeneities in the flow regimes – the most convincing to date – implied that there is a systematic process responsible for the discrepancy revealed in the previous models. Here the surface water-groundwater interaction is considered the most plausible explanation. The study by Wolski and Savenije (2004) provided some insight into this process. Since surface water-groundwater interactions were not simulated explicitly in the previous models, it was hypothesised that a proper incorporation of this process would allow for simulation of the observed behaviour of the system, without the introduction of additional parameters to correct for the non-homogeneities.

4.2 Reservoir model

The model used in this study is similar in structure to the previous models prepared for the Okavango Delta (Denčer et al., 1987; SMEC, 1990; Scudder et al., 1993; Gieske, 1997; WTC, 1997). The Okavango Delta is represented as a set of inter-linked linear reservoirs. For each reservoir, the volume of surface water $V$ is calculated on a monthly basis, according to the equation:

$$\frac{dV}{dt} = I + E - P - Q - Q_{\text{inf}}$$

where $I$ is surface water inflow (from upstream reservoir or from the feeding Okavango River), $E$ is the evaporation from the inundated area, $P$ is the rainfall over the inundated area, and $Q_{\text{inf}}$ is the infiltration to groundwater. All units in (Mm$^3$-month$^{-1}$). The outflow $Q$ from each of the reservoirs is related to the volume of water stored above a certain threshold volume $V_{\text{thres}}$, corresponding to the outlet level in a simple linear manner, i.e.
\[ Q = \frac{1}{k} \cdot (V - V_{\text{dss}}) \]  

where \( k \), expressed in units of time (months), is known as the time constant of a reservoir. It is a measure for the resistance to flow in the system, and directly expresses mean residence time. To represent the relation between the inundated area \( A \) that is associated with surface water volume \( V \), a simple power relation is used:

\[ A = (n \cdot V)^b \]  

where the coefficients \( n \) and \( b \) are determined on the basis of topographical data.

**Basic reservoir setting**

Our model consists of 8 units representing the major distributaries of the Delta. Initial work on the model included as much as 23 units. With this subdivision a better representation was expected of the spatial heterogeneity in the flow system properties. However, the preliminary runs showed that the detailed model gained little in terms of performance, and appeared to be over-parameterised. Hence the number of units was reduced. Based on similarities and differences in flooding behaviour the units were rearranged into a smaller number. This process also provided insight into the linkages between the different distributaries of the Delta.

**Reservoir links**

An upstream reservoir has separate outlets feeding downstream reservoirs. More than one outlet can provide a link between two reservoirs. In this way the non-linear character of the flow could be simulated. Whether or not an additional outlet was implemented for a given reservoir link, was decided based on the analyses of the dynamics of the flood units downstream. This way of linking reservoirs is different from the previous models, where surface water reservoirs had only one outlet, and a fixed proportion of outflow was distributed to downstream units.

**Surface water-groundwater interactions**

In the old models, infiltration was either not explicitly represented (Gieske, 1997), or it was represented by a simple function relating it to the volume stored in the surface water reservoir (Dincer et al., 1987), or it was taken as an amount depending on the increase of the inundated area (SMEC, 1990; Scudder et al., 1993). In the WTC (1997) model, it was simulated using a function accounting for the duration of the no-flood period preceding the inundation. In all these models the infiltrated water was no longer accounted for in the model.
In our model, infiltration and groundwater flow processes are explicitly represented by a set of reservoirs. Each of the surface water units has been divided into five sub-units represented by independent groundwater reservoir sets, comprising two reservoirs each: a floodplain groundwater reservoir of area $A_F$ and an island groundwater reservoir of area $A_i$. Each floodplain groundwater reservoir corresponds to an area flooded under a progressively increasing flood, i.e., the first one represents a floodplain flooded by, for example, 250 km$^2$, the second one the additional flooded area when the flood extent increases to 500 km$^2$ etc. Infiltration $Q_{in}$ fills the empty storage in the groundwater reservoir immediately after the flood extent has exceeded the threshold flood size. At a monthly time step, this is consistent with observations (Wolski and Savenije, 2004). Subsequently, groundwater starts flowing from the floodplain to the islands, as a function of the gradient between the floodplain and island groundwater reservoirs. The groundwater flow $Q_{gw}$ is a function of the difference between the groundwater levels in the floodplain ($h_F$) and on the island ($h_i$), i.e.:

$$Q_{gw} = T \cdot (h_F - h_i)$$  \hspace{1cm} (4)

where $T$ is the transmissivity (m$^2$ month$^{-1}$) of the floodplain-island system.

The water balance equation for a floodplain groundwater reservoir hence reads:

$$\frac{dV_F}{dt} = P - E + Q_{in} - Q_{gw}$$ \hspace{1cm} (5)

and for the island groundwater reservoir:

$$\frac{dV_i}{dt} = P - E + Q_{gw}$$ \hspace{1cm} (6)

The water level in a groundwater reservoir is using following equation:

$$h = \frac{V}{A \cdot s_y}$$ \hspace{1cm} (7)

where $V$ is the volume stored in a reservoir above a certain reference level, $A$ is the area of the reservoir and $s_y$ is the coefficient expressing the specific yield of the aquifer.

If the floodplain groundwater reservoir is covered by flood water, it is replenished after each time step. If it is no longer covered, the storage continues to deplete by the flow towards the island and by

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evaporation. The island groundwater is emptied by evaporation, which maintains the floodplain-island groundwater flow through Eq. (4).

**Evaporation**

Evaporation from all units in the model is calculated using the concept of potential evaporation ($E_p$):

$$ E = e_f \cdot E_p \cdot A $$

(8)

where $e_f$ is a coefficient reflecting availability of water, and $A$ is the area of the unit. $E_p$ is expressed in (m-month$^{-1}$). For a surface water reservoir $e_f$ is expressed as:

$$ e_f = \frac{A_s}{A} $$

(9)

where $A_s$ is the inundated area. For a floodplain groundwater reservoir (if not inundated) and island groundwater reservoir it is taken as:

$$ e_f = \max \left( 1 - \frac{d}{d_{\text{ext}}}, 0 \right) $$

(10)

where $d_{\text{ext}}$ is the so-called extinction depth, and $d$ is the actual depth to groundwater table, which is calculated from the volume $V_r$, analogous to Eq. (7). Eq. (10) demonstrates a linear decrease of evaporation and transpiration with groundwater depth. Evaporation from groundwater is commonly expressed in this way in generic groundwater models (e.g. Chiang and Kinzelbach, 2001).

To obtain the potential evaporation rate for the Okavango Delta, data from several sources were analysed. Some short-term measurements of meteorological conditions are available from the Delta proper. However, the time series of meteorological measurements at Maun station (on the fringe of the Delta) is available since 1970. It was therefore logical to relate potential evaporation conditions in the Delta proper to the conditions at that station. The Penman-Monteith combination equation (Allen et al., 1998) was used to calculate the reference crop transpiration $E_{\text{ref}}$ for Maun and for a station in the Delta proper, Nxaraga, for which data are available since 1999. The comparison of these rates (see Fig. 3) shows that in the Delta rates are lower than at Maun, by approximately 0.5 mm·d$^{-1}$. This difference is caused by the higher humidity and the lower temperatures in the Delta compared to Maun, which, considering the "oasis" character of the Delta, is not unexpected. The limited evaporation and transpiration data from an eddy covariance system, located at an inundated papyrus site and at an inundated typical seasonal floodplain, show that actual evaporation rates ($E_{\text{act}}$) from these inundated sites are lower than the reference rates at Maun, and that the ratio varies in time (Fig. 3). The ratio of $E_{\text{act}}/E_{\text{ref}} = 0.8$ in summer (February) and 0.6 in winter (June). This difference results from the combined effect of microclimatic conditions in the Delta, the shading of water surface by plants, and of the plant growth cycle, where growth occurs only in summer and plants are dormant during the wintertime. These results agree with the results obtained from analyses of the isotopic
composition of flood water (Hutton and Dinçer, 1976), revealing that the winter-time evaporation from inundated areas is dominated by open water evaporation, while during summertime, the contribution of transpiration and open water evaporation are approximately equal. In view of these facts, an empirical time-varying coefficient has been introduced relating the potential evaporation rate \( E_p \) to Maun's \( E_{ref} \). This approach is similar in nature to the commonly used \( k_p \) factor. The coefficient used varies between 0.65 for winter-time (June-July) to 0.85 for summertime (December/January).

**Rainfall**
Rainfall input was calculated according to the following equation:

\[
P = p \cdot A
\]

(11)

where \( p \) is the rainfall rate in \( \text{m-month}^{-1} \) and \( A \) is the area. In case of a surface water reservoir \( A \) is the inundated area \( A_S \); in case of a floodplain groundwater reservoir it is the area that remains un-inundated; in case of an island groundwater reservoir, it is the area \( A_I \) of that reservoir. The rainfall rate \( p \) was calculated separately for each of the surface water units as an inverse distance-weighted average of uncorrected rainfall from the two long-term rainfall stations, Maun and Shakawe.

**Calculation procedure**
Eqs. (1), (5) and (6) are solved iteratively for each of the model units. This approach has the advantage of being exact, i.e. not being dependent on the length of the computational time step, and it allows the reservoir constant \( k \) to be smaller than the computational time step. The model has been coded as an MS Excel macro, and coding was verified by thoroughly checking the closures of water balances of all model units.

**Derivation of model parameters**
The model operates with a set of parameters for each of the computational units. These are:
- area parameters: \( A_r \) and \( A_i \)
- topography parameters: \( b \) and \( n \)
- evaporation parameters: \( d_{ref}, d_{act} \)
- flow parameters: \( T, V_{lives} \) and \( k \)

Area parameters have been derived from the available satellite images and land cover maps (McCarthy and Gumbricht, 2004). The floodplain reservoir area was mapped as the area of both active and potential floodplains. For each of the model units, the area of the floodplain groundwater reservoirs sums up to the surface water unit area. The island area has been determined as the area of woody species-covered islands not further than 1000 m away from the floodplain-island boundary.
This was done since the low groundwater gradients, and the effect of transpiration uptake of groundwater by trees, limit the area of influence of the flooding to approximately 1000 m from the floodplain-island boundary (Wolski and Savenije, 2004).

The topographic parameters \( n \) and \( b \) have been derived from the available Digital Elevation Model (DEM) (Gumbricht et al., 2004a). Flood extent and surface storage volumes have been calculated for each of the units by intersecting the unit with the DEM. Not surprisingly, the coefficients obtained from this procedure indicated a much "flatter" surface area-volume relationship than obtained from field measurements and implemented in previous models (Fig. 4). This is due to different hydrological behaviour of the system at the scales of a single floodplain and an entire distributary unit. For a small floodplain, the initial increase in stored volume occurs as a rapid increase in area; at higher water levels, volume increases in the "bowl-shaped" morphology of the floodplain result in increases in water depth without much increase in flooded area. At the scale of the model unit, however, the increase in stored volume is accommodated by an increase in the number of floodplains flooded; that is, by lateral expansion of the flooded area, and thus the effect of the bowl-shaped topography of the individual floodplain is not manifest.

The evaporation parameters \( d_{\text{ext}} \) were estimated based on the observed behaviour of the groundwater table (Wolski and Savenije, 2004). For a floodplain groundwater reservoir \( d_{\text{ext,F}} \) was taken as 5 m and for an island groundwater reservoir \( d_{\text{ext,I}} \) as 20 m.

Flow parameters, due to their conceptual character, were calibrated in the model. In the light of the low sensitivity of the model to changes of \( T \), no spatial variation in the transmissivity has been introduced.

**Calibration of the model**

Calibration of the model targeted the most reliable measurements of system response, namely: a) discharges observed at the outlet of the Boro distributary at Maun, and b) flood sizes observed for each of the units, as well as for the entire system, with the Boro distributary being the primary target. Discharges observed at other outlets (e.g. Xudum at Toteng), which are not very reliable due to frequent gaps in data series, were considered only as supplementary data. The hydrological monitoring network of the Okavango Delta comprises over 30 water level and discharge measurement sites, all of which are located exclusively in channels. As stipulated in the result section, there are considerable differences in magnitude and dynamics between channel discharges (which are monitored) and distributary discharges (which are simulated in the model). Hydrological behaviour observed at the monitoring sites inside the Delta proper were therefore not considered during model calibration. Calibration was done by trial-and-error adjustments of model parameters.
4.3. GIS model

In order to translate the lumped flood extent obtained from the reservoir model into a flood distribution map, the available series of flood maps obtained from classification of satellite images have been used. Analysis of the maps reveals that within the distributaries, floods of similar size have rather similar spatial distribution, although some variation occurs. That variation partly result from image processing artefacts, i.e. misregistration and misclassification of NOAA-AVHRR images, from which the flood maps were derived. Part of that variation is probably related to the natural variation in flood distribution resulting from either secular structural changes within the distributary, or from effects of vegetation-hydrology feedbacks or non-linearity of the system. Since with the available data set it was not possible to differentiate between these causes, we decided to treat all of the within-distributary variation as an apparently random variation, and use the stochastic approach in constructing flood maps for given flood size. Main elements of the procedure are presented below.

Each pixel appears either flooded or not flooded in the flood maps. In a stationary, deterministic situation, for each of the pixels there would be a flood of size a that is the “smallest inundating” flood for that pixel, i.e. floods of smaller size will not inundate that pixel, while floods of larger size do. Since there is variation in flood distribution, there is no sharp distinction between inundating and non-inundating floods. A pixel may appear not inundated by a certain flood, while it has been inundated by smaller floods. Thus, for each pixel, floods of different sizes x have an associated probability f(x) of being the “smallest inundating” flood. That probability can be described by a probability density function, and we assume it to be the normal distribution, i.e.:

\[ f(x) = \frac{1}{\sqrt{2\pi}} e^{-\frac{x^2}{2}} \]  \hspace{1cm} (12)

\[ z = \frac{x - \mu}{\sigma} \]

where \( z \) is a standard normal variable and \( \mu \) and \( \sigma \) are parameters.

The corresponding standard normal probability distribution function has the form:

\[ F(z) = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{z} e^{-u^2/2} du \]  \hspace{1cm} (13)

where \( u \) is the dummy variable of integration. The function expressed by eq. (13) has no analytical solution, but a polynomial approximation is available (Chow et al., 1988). \( F(z) \) denotes the probability that a given pixel is inundated by flood of size \( x \), i.e.

\[ P(x) = F(z) \]  \hspace{1cm} (14)
while not being flooded by that flood has the probability of:

\[ P(x) = 1 - F(z) \]  
\[(15)\]

A series of independent events, as is in this context the series of n flood maps, has therefore a joint likelihood of occurrence:

\[ L = P(x_1) \cdot P(x_2) \cdot \ldots \cdot P(x_n) \]  
\[(16)\]

where \( P(x_i) \) is expressed by eq.(14) in case the pixel is inundated, or by eq.(15) if it is not.

Parameters describing the probability density function (Eq. 12), i.e. \( \mu \) and \( \sigma \), can be determined for each of the pixels using the method of maximum likelihood, i.e. by maximizing the value of the joint likelihood of occurrence \( L \) expressed by Eq.(16).

Once \( \mu \) and \( \sigma \) have been determined on a pixel-by-pixel basis, maps depicting the probability of being inundated by a flood of size \( x \) can be produced based on solution to Eq.(13). The output of the reservoir model, i.e. the flood extent in a given computational unit at the end of each computational time step, can therefore be presented in the form of a probability of inundation map. Composite products such as flood frequency and flood duration maps can subsequently be obtained with associated probabilities.
5. Results and discussion

Outflows
Model performance criteria are summarized in Table 1. In terms of outflow from the Boro, the correlation coefficient of the observed and simulated monthly flows is 0.91 and a MSE of 11.9 Mm$^3$. These results are better than those obtained by the old models that used correction factors, and similar to those obtained by Gieske (1997). In terms of Boro flow duration and flood size exceedance, the model performs very well (Fig. 5, Fig. 6). Simulation of the Toteng discharge is somewhat worse (Fig. 5), but the general pattern of wet and dry years is well represented.

Flood size
The distributaries that display strong seasonal variation in flooding are well represented by the model (Thaoge, Xudum and Boro in Fig. 7). Both maximum annual flood extent and timing of the Okavango river flood are well replicated by the model. For the eastern distributaries, which display less seasonal variation, model performance is poorer with discrepancies occurring during the rainy seasons (Mboroga in Fig. 7). The model tends to overestimate the flood extent in response to rainfall, while observations indicate a decline of the flood extent during most rainy seasons. The problem probably results from the difference in flood expansion in response to rainfall (which spreads instantaneously) as opposed to the arrival of a flood wave (which slowly propagates). Rainfall produces an increase in the floodplain water volume that is uniform, equally distributed over the inundated area, while the Okavango River flood wave produces a wedge-like volume distribution. The latter, due to higher water level gradients as compared to the former, results in stronger expansion of the flood. Inability to simulate that is, unfortunately, one of the limitations of our model. Nonetheless, the model is able to reproduce the dynamics of the flooding in the eastern distributaries similar to the observed, especially when contrasted with the dynamics of the central and western distributaries.

Boro flow regimes
The model simulates the 1974-1981 Boro outflows well. Underestimation of outflows is observed only during the last year of the period. As simulated by the model, the effects of increased outflows during the “high flow regime” period are caused by the nature of surface water-groundwater interactions and by the extremely high (1170 mm·a$^{-1}$) rainfall of the 1973/74 rainy season. That rainfall caused recharge of the floodplain and island groundwater (Fig. 10). As a result, the infiltration was reduced, and more water was available for outflow to the next cell. Groundwater levels remained high during the 1974-1981 period, and were supplemented by rainfall recharge during 1976 and 1979. In our model, the “high flow regime” effect is visible throughout the Delta, and not only in the Boro system.
To fully verify this explanation of the Boro "high flow regime" groundwater levels data are necessary. These, however, do not exist.

**Shifts & changes in the system**
In recent years, an increase in the flood extent in the Xudum has been observed, as compared to the neighbouring Boro. This is also visible in the model results: the model consistently underestimates the flood extent in the Xudum after 1995. This situation is similar to the Boro "high flow regime" anomaly, encountered by the authors of earlier models. The Boro flow anomaly, however, could be well simulated by surface water-groundwater interaction and a sequence of high rainfall years, whereas the same does not explain the Xudum anomaly. This shift may therefore be the effect of a change in system properties, with more water flowing towards Xudum as compared to other distributaries. It has been earlier suggested by Gumbricht et al. (2004b), that the Xudum is gaining at the expense of Boro. However, the results of the model suggest that the Xudum gains water at the expense of the Thaoge, as the model seems to overestimate flood sizes in that distributary after 1995, particularly during 1996-1999 period (Fig. 7). The shift has dramatic socio-economic and ecological consequences, and it is therefore important to understand its causes. However, at this stage this hypothesis cannot yet be verified. For the time being, we shall treat this phenomenon as an unsystematic change in the system's properties, but will seek the explanation through further research before accepting it as a shift in a model parameter.

**Active and passive storage**
During the calibration of the model, it appeared that the flood extent in the units representing the central Delta (Nqoga and Maunachira in Fig. 2), could not be reproduced properly by the model. The modelled annual peak occurred much earlier than the observed. The comparison of the observed water levels and the observed flood extent time series revealed that these were clearly out of phase (Fig. 8). Although the modelled volume showed a similar pattern as the observed water levels, they were not corresponding to the observed flood extent. To deal with this discrepancy, we introduced active and passive storage within each of these units. Active storage is the one that is replenished by inflow from upstream and which determines the outflow to the downstream reservoirs. Passive storage receives water as a spill-over from active storage. The variation in the amount of water in the passive storage is mainly responsible for the flood extent in the unit. This distinction agrees with reality, where passive storage is mostly shallow floodplain storage that does not contribute to flow, but has a large flood extent, whereas the active storage is part of the conveyance network. By introducing this distinction into the model, the flood extent pattern of the central units was reproduced better, while the dynamics of the downstream units was maintained.
Spatially distributed flood characteristics
The results of hydrological model are summarised as a frequency curve for annual inundation for the 1985-2000 period in Fig. 9a. These results are compared to the “observed” inundation frequency obtained from flood maps, depicting the maximum annual flood extent for each of the years in that period (Fig. 9b and c). Differences occur mainly in the Mboroga system, where the combined model overestimates the frequency of inundation, and this reflects the difficulty of the hydrological model to represent the effects of rain-induced flooding. The suggested shift in flow distribution between the Thaoge and the Xudum, which was not accounted for in the hydrological model, is visible through an underestimation of the inundation frequency in the Xudum and an overestimation in the Thaoge.

Flood frequency maps and other products of the integrated hydrological-GIS model can be used for assessing effects of upstream development and climate change scenarios on the eco-system of the Okavango Delta (Murray-Hudson et al., this issue).

Channel flows vs distributary flows
In the model, there is no differentiation between channel and floodplain discharges. Rather, the model simulates the sum of these two fluxes across unit boundaries as distributary flows. Due to the size of the model units, a comparison between simulated and measured channel discharges is only possible at two sites (Fig. 11). This figure reveals that the magnitude and seasonal dynamics of channel flow differ considerably from that of the entire distributary (which includes the flow through the floodplain).

At both sites, the majority of the annual flood volume passes through the floodplain system. This fact has important implications for understanding the role of channel sedimentation and blockage in large-scale processes such as the abandonment or reopening of a distributary (e.g. McCarthy et al., 1988, McCarthy and Ellery, 1997). It also implies that we need to re-design the hydrological monitoring network.

Water balances
Traditionally, the water balance of the Okavango Delta was lumped for the entire delta with evaporation calculated as a rest term, representing one “bulk” term for evaporation and transpiration. This model allows the calculation of a system water balance with evaporation separated into different components, namely open water evaporation, non-inundated floodplain evaporation and island evaporation (Table 2). It appears that evaporation from the open water surface accounts for approximately 58% of the total evaporation, while that from the non-inundated floodplains for approximately 17%. An important element is the island evaporation: it is responsible for approximately 25% of the total evaporation, 50% of which is provided by local rainfall. The remainder is supplied by groundwater inflow from the floodplains.
6. Summary and conclusions

The Okavango Delta system can be adequately described by a conceptual model consisting of only 9 units. The conceptual model is straightforward with a minimum number of parameters to prevent equifinality problems. Many of the parameters are physically based and can be obtained through independent measurements.

Previous models were not capable of representing long term variability in the time-series, because the main mechanism for over year storage, the interaction with the groundwater and the islands, was not present. Key to the good performance of the model is the surface water-groundwater interaction and the use of multiple links with threshold values between cells. Also local rainfall has an important role to play, since it affects the storage availability in the floodplains and the islands.

The GIS model that translates the flood extent in the hydrological model to inundation maps is a very useful tool to translate hydrological scenarios into eco-system response functions and to study the ecological impacts of different upstream development scenarios.
Tables

Table 1 Model performance criteria

<table>
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<tr>
<th></th>
<th>This model</th>
<th>SMEC</th>
<th>IUCN</th>
<th>Gieske***</th>
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<tr>
<td></td>
<td>MSE</td>
<td>r</td>
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<td>monthly</td>
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* with three parameter sets; ** with rainfall adjustment; *** with CRD
Table 2 Long term (January 1969-December 2002) water balance of the Okavango Delta

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<th>M m³ a⁻¹</th>
<th>mm a⁻¹</th>
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<td>Floodplains (9327 km²)</td>
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<td>469</td>
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<td>Islands (4090 km²)</td>
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<td>463</td>
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<td><strong>Total inputs</strong></td>
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<td></td>
<td>100</td>
</tr>
<tr>
<td><strong>Outputs</strong></td>
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<tr>
<td>Outflow*</td>
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<td>2 (of total output)</td>
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<tr>
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</tr>
<tr>
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<td><strong>Change in storage</strong></td>
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* Maun and Toteng only  ** calculated over time-varying area
Figures

Fig. 1 Okavango Delta and its main features
Fig. 2 Basic model setting a) units represented by surface water reservoirs, b) schematic for unit reservoir setting (SW- surface water reservoir, FGW – floodplain groundwater reservoir, IGW – island groundwater reservoir)
Fig. 3 Comparison of $e_{\text{ref}}$ for Maun and Nxaraga and $e_{\text{sec}}$ from an eddy covariance system at papyrus and seasonal floodplain site (location of all the sites in Fig. 1)
Fig. 4 Volume-area curves for a small floodplain, used in the previous models and derived from DEM for large units.
Fig. 5 Observed and simulated discharges at Maun and Toteng.
Fig. 6 Observed and simulated a) flow duration at Maun and b) flood size exceedance in Boro distributary.
Fig. 7 Observed and simulated flood sizes in Boro, Xudum, Thaoge and Mboroga distributaries.
Fig. 8 Observed water levels and flood extent for the Nqoga unit
Fig. 9 Flood frequency map for the 1990-2000 period (observed, simulated and difference)
Fig. 10 Island groundwater storage ($V_I$) and floodplain-island groundwater flow ($Q_{gw}$) for Boro distributary as simulated by the model
Fig. 11 Comparison of observed channel flows and modelled distributary flows for two sites: Mmadinare and Gaenga (location in Fig. 1)


review of the Southern Okavango Integrated Water Development Project. IUCN, Gland, Switzerland.

SMEC (Snowy Mountains Engineering Corporation), 1990. Southern Okavango Integrated Water Development Project, Department of Water Affairs, Gaborone, Botswana.


