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REPORT

Modelling water-table depth in a primary aquifer to identify potential wetland hydrogeomorphic settings on the northern Maputaland Coastal Plain, KwaZulu-Natal, South Africa

Bruce E. Kelbe[1] **· Althea T. Grundling**[2,3,4] **· Jonathan S. Price**[3]

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Abstract The primary aquifer on the Maputaland Coastal Plain in northern KwaZulu-Natal, South Africa, is the principal source of water for rivers, lakes and most of the wetlands in dry periods, and is recharged by these systems in wet periods. Modelling hydrologic conditions that control regional water-table depth can provide insight into the spatial patterns of wetland occurrence and of the persistence of wet conditions that control their character. This project used a groundwater model (MODFLOW) to simulate 10-year water-table fluctuations on the Maputaland Coastal Plain from January 2000 to December 2010, to contrast the conditions between wet and dry years. Remote sensing imagery was used to map “permanent” and “temporary” wetlands in dry and wet years to evaluate the effectiveness of identifying the suitable conditions for their formation using numerical modelling techniques. The results confirm that topography plays an important role on a sub-regional and local level to support wetland formation. The wetlands’ extent and distribution are directly associated with the spatial and temporal variations of the water table in relation to the topographical profile. Groundwater discharge zones in the lowland (1–50 masl) areas support more permanent wetlands with dominantly peat or high organic soil substrates, including swamp forest and most of the permanent open water areas. Most temporary wetlands associated with low-percentage clay occurrence are through-flow low-lying interdune systems characterised by regional fluctuation of the water table, while other temporary wetlands are perched or partially perched. The latter requires a more sophisticated saturated-unsaturated modelling approach.

Keywords South Africa · Primary aquifer · Wetlands · Numerical modelling

✉ Bruce E. Kelbe
bekelbe@gmail.com

1 Hydrological Research Unit, University of Zululand, Private Bag X1001, KwaDlangezwa, KZN 3886, South Africa

2 Agricultural Research Council - Institute for Soil, Climate and Water, Private Bag X79, Pretoria, Gauteng 0001, South Africa

3 Department of Geography and Environmental Management, University of Waterloo, 200 University Avenue West, Waterloo, ON, Canada

4 Applied Behavioural Ecology and Ecosystem Research Unit, University of South Africa (UNISA), Pretoria, South Africa

Introduction

Wetlands form where water is present at or near the land surface for a sufficiently long time to promote hydric soils and support vegetation communities adapted to wet conditions. These conditions can arise when the hydrogeomorphic setting and climate result in a high water table connected to the regional groundwater regime, or where perched water tables intersect the surface topography. The Maputaland Coastal Plain, also known as the Mozambique Coastal Plain, in the northeastern region of KwaZulu-Natal Province in South Africa, consists of a low relief, undulating sandy dune landscape that contains the highest percentage of wetland area per province area in South Africa (SANBI 2010) and 60 % of South Africa’s known peatlands (Grundling et al. 1998). Rainwater infiltrates into the coastal dunes to recharge the shallow aquifer linked to groundwater-dependent ecosystems (Taylor 1991). Many interdune or topographic lows in the vicinity of high a water table are wet (inundated or saturated), forming aquifer-dependent ecosystems of the Maputaland Coastal Plain (Colvin et al. 2007; Taylor et al. 2006).

In this area, the shallow aquifer is the dominant hydrological feature that is closely linked to the aquatic and terrestrial ecology (Taylor et al. 2006; Colvin et al. 2007; Le Maitre and Colvin 2008; Kelbe and Germishuyse 2010). There is a need to clarify the source and persistence of water in these wetlands (Begg 1989), since this affects wetland form and function (Barker and Maltby 2009). While this information is not generally available for wetlands of the Maputaland Coastal Plain (Ewart-Smith et al. 2006), there is good evidence that many wetlands serve as groundwater discharge areas (Taylor et al. 2006). The wetlands may be linked through surface drainage systems forming low gradient streams that can directly influence the water-table profile and, hence, the wetlands themselves. However, wetlands can form from perched water tables that are not directly connected to the regional water table (Dempster et al. 2006). It is difficult to identify these linked and perched systems without detailed field studies. A regional groundwater model that extrapolates the depth to water table based on the regional topography, geomorphology, climatology and lithology, coupled with information on wetland distribution and character, will assist in understanding the nature of wetland distribution and type, as well as their susceptibility to land-use and climate changes.

Determination of the spatial patterns of water-table depth is often derived from the interpolation and extrapolation of water level measurements at monitoring sites and exposed water surfaces (Kulasova et al. 2014) that are assumed to be extensions of the groundwater system. Often these interpolation methods fail to include the impact of changing groundwater fluxes associated with known hydrogeomorphic features including drainage boundaries, topographic expression and lithological discontinuities or heterogeneity. These fluxes can induce significant changes in the temporal and spatial patterns of groundwater storage that can produce rapid changes in the water-table profile, and hence the requisite conditions for wetland development. This is particularly relevant in the shallow aquifers along the coastal plain where the water-table hydrographs can resemble the stream hydrographs (Kelbe and Germishuyse 2010). Groundwater models can be used to characterise the spatial and temporal patterns of groundwater storage (Gilvear and Bradley 2009) that are linked to the distribution and function of wetlands and lakes (Winter 1999; Kelbe and Germishuyse 2000).

The application of numerical methods to support environmental studies is a pragmatic approach that provides increasingly reliable estimates of the form, function and dynamics of aquatic systems as conceptual modelling and data assimilation of the system progresses during model development and calibration. Hendricks Franssen et al. (2008) stated that remote sensing data are useful for groundwater modelling in large semi-arid areas where there is a lack of hydrological data. Groundwater modelling and remote sensing applications are important tools for managing wetlands and understanding their functioning in the landscape (Trigg et al. 2014). There have been several investigations on the application of remote sensing to support groundwater modelling in regions of limited accessibility for ground data acquisition in southern Africa (Brunner et al. 2004, 2007, 2008). However, there is still uncertainty in parameter estimation in the application of remote sensing (Jovanovik et al. 2014). If the appropriate conceptual models and ground truth information are available to support the simulation of all the relevant driving features of the system that create the water level responses, the numerical model will provide a strong analytical tool to evaluate the groundwater relations driving the environmental system.

Insight is needed into how groundwater discharge and the depth to water table relate to wetland types of the Maputaland Coastal Plain. The derivation of a reliable estimate of the water-table profile and its variability are important factors in the study of these environmental systems, particularly the distribution of the permanent and temporary wetlands. An accurate map of water-table distributions in these situations would greatly assist in the identification of wetland types and their dependency on the regional aquifer. Various studies have been published on the use of groundwater models in support of environmental studies for the study region (Kelbe and Germishuyse 2000, 2001, 2010; Været et al. 2009). For example Eucalyptus plantations (phreatophytic trees) are known for their high evapotranspiration rates and deep roots, lowering the groundwater table by more than 20 m (Scott et al. 2000; Brites 2013; Benyon et al. 2006; Canadell et al. 1996) and have been identified as an emerging threat to the wetlands on the Maputaland Coastal Plain (Rawlins 1991; Walters et al. 2011). Macfarlane et al. (2012) and Grundling et al. (2013a) reported an increase in afforestation in the study area. Dennis (2014) stated that forestry affects the inflows and water levels in the lakes and recommended that no forestry plantations be established within the recharge zone of sensitive wetlands, as this significantly increases the water deficit and potentially impacts these groundwater-dependent ecosystems. The groundwater model would greatly assist in identifying these recharge zones and their potential impacts from anthropogenic changes in land-use.

The unique importance of groundwater in this area for the estuarine ecosystems during severe droughts has been studied by Taylor et al. (2006). These studies and subsequent changes in land-use have led to management controls that enhance groundwater recharge to protect the ecological resources of the Maputaland Coastal Plain. The aim of this study is to use groundwater modelling to derive the best estimate of the regional spatio-temporal distribution of the water table during a wet and dry period, to aid in the delineation and characterisation of wetland types in regions with limited ground truth data.

Study area

Location

The Maputaland Coastal Plain in north-eastern KwaZulu-Natal Province, South Africa (Fig. 1) is renowned for its biodiversity, conservation areas and World Heritage Site that include a variety of fresh and saline water wetlands such as swamp forest, saline reed swamp, salt marsh, submerged macrophyte beds, mangroves and riverine woodlands (Taylor 1991). The study area is situated in the north-eastern part of the Maputaland Coastal Plain between the Tembe Elephant Park and the Kosi Bay lake system (Fig. 1a). Economic activity on the Maputaland Coastal Plain consists predominantly of subsistence agriculture (croplands and rangelands), forestry (plantations) and eco-tourism centred around the coastal wetlands (Fig. 1c). The iSimangaliso Wetland Park is a World Heritage Site that protects the

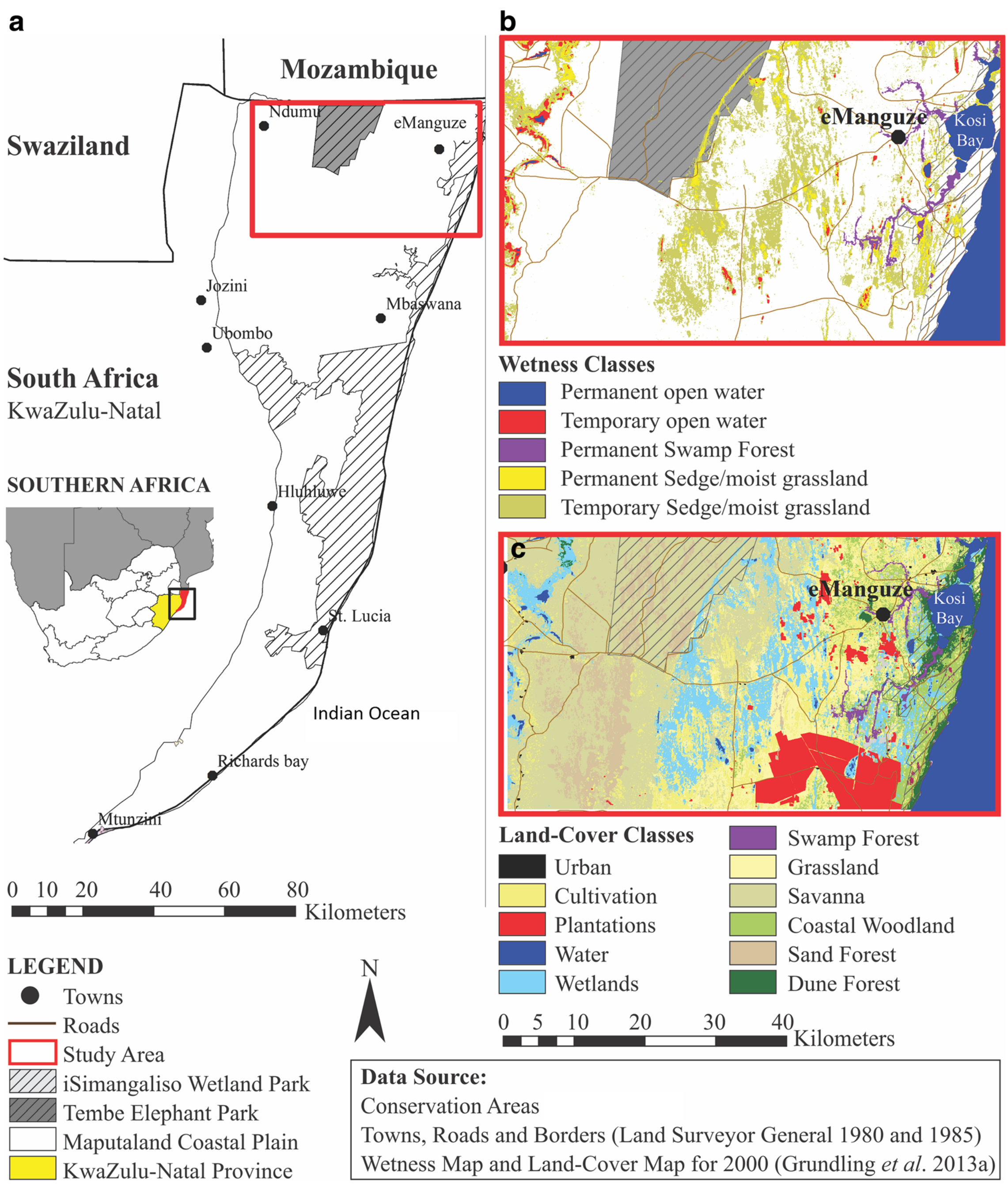

Fig. 1 **a** The study area on the Maputaland Coastal Plain in north-eastern KwaZulu-Natal Province, South Africa (Grundling et al. 2013a), **b** The distribution of wetlands and **c** Land-cover classes (Grundling et al. 2013a). Data Source: Conservation Areas (SANBI 2010), Town, Road and Border (Land Surveyor General 1980 and 1985), Wetness Map and Land-Cover Map for 2000 (Grundling et al. 2013a)

environment along the coastal strip around the Kosi Bay lake system up to the Mozambique border. The Tembe Elephant Park is a proclaimed community conservation area that is being linked to the Maputo Elephant Park as part of a Transfrontier park with Mozambique and Swaziland. The Maputaland Coastal Plain has poor soils (Lubke et al. 1996) that are generally unsuitable for commercial grain farming. However, the region is under severe threat from regulated large-scale commercial forestry, as well as an increase in uncontrolled small-scale forestry by subsistence farmers, both of which can have significant impacts on groundwater levels and wetlands (Været et al. 2009; Grundling et al. 2013a; Brites 2013). The local communities in the region rely on subsistence agriculture in wetlands for crop production.

Wetlands

Grundling et al. (2013a) used Landsat TM and ETM imagery acquired for 1992 and 2008 (dry years) and Landsat ETM for 2000 (wet year) along with ancillary data such as a digital elevation model (DEM), vegetation and soil maps to identify and map permanent and temporary (inland) wetlands and open water (Fig. 1b), based on land-cover classification for the different years. All three datasets were used for land-cover change analysis to describe the spatial extent and distribution of wetlands and open water as well as land-use classes during the three different years to determine wetland loss from land-use changes due to cultivation, plantation and urbanisation.

The study area hosts a complex array of wetland types that range from "permanent wetlands" with dominantly peat or high organic soil substrates to "temporary wetlands" with mineral soils (Grundling et al. 2013a; Pretorius 2011). The extent of wetlands varies in response to periods of water surplus or drought, from large temporary wetland systems to permanent linear interdune wetlands between the parabolic dunes (KwaMbonambi Formation).

Permanent wetlands have a relatively fixed boundary, e.g., peat swamp forests (Grobler 2009), while sedge/moist grassland wetlands that occur on the deep sandy soil in areas where the water-table fluctuations are greater (conditions not ideal for peat development) are referred to as temporary wetlands (Pretorius 2011). The boundaries of temporary wetlands appear to grow or shrink in wet or dry periods (Begg 1989), potentially causing their area to be underestimated in periods of water shortage (Grundling et al. 2013a). During very wet years, some areas including wetlands can be temporarily inundated with pools of open water for a short period. These can be described as "temporary open water" (Grundling et al. 2013a). In contrast, there are "permanent open water" areas including the Kosi Bay lake system and smaller lakes such as Lake Shengeza (Grundling et al. 2013a).

In general the regional geology slopes towards the east and the precipitation (rainfall) gradient decreases from east (>820 mm) to west (~680 mm; Grundling et al. 2013a). The upland (≥50 masl) has a greater proportion of temporary sedge/moist grassland wetlands, while the lowland areas (<50 masl), mostly in the east, host most of the permanent wetlands, including swamp forest, as well as some temporary wetlands and most of the permanent open water (Figs. 1b and 2a). Groundwater recharge takes place when there is sufficient rainfall, while groundwater discharge occurs through drainage lines in low-lying areas, facilitated by the underlying regional geology that slopes towards the east. Consequently, the permanent open water areas (Kosi Bay lake system and Lake Shengeza), which represent 2–3 % of the total study area, and all of the swamp forest are congruent with the high water table in the coastal region.

The wetland distribution and temporal character are related to the nature of the aquifer, topography and rainfall distribution, i.e., hydrogeomorphic setting (Grundling et al. 2013a). Hydrogeomorphic wetland units identified by Grundling et al. (2014) include: a floodplain, channelled valley-bottom, unchannelled valley-bottom, depressions and seep areas. However, wetland occurrence is not dependent on rainfall or elevation but rather depth to water table, which is dependent on the hydraulic characteristics of the regional aquifer and localised topographical features and associated hydrological processes. Some temporary wetlands are perched pans (e.g., Kwamsomi Pan, parallel to the Muzi wetland system) while others are flow-through interdune systems (Grundling 2014).

Hydrogeology

The Maputaland Coastal Plain was formed by sedimentary processes during periods of marine regressions and transgressions (Botha et al. 2013) that created a sedimentary sequence of unconsolidated formations. Subsequent aeolian depositions formed paleo-dune ridges orientated parallel to the coast and more recent high frontal dunes along the shoreline (Fig. 2c). The coastal plain is characterised by a sequence of sediments overlying consolidated rocks of Jurassic basalts and rhyolitic rocks that generally slope to the east at an angle of about 3° (Botha et al. 2013). During the Cretaceous Period much of the area was below sea level, creating a hydrogeological unit of claystones and siltstones with very low hydraulic conductivity, porosity and storativity, which behaves as an aquiclude with residual brackish water (Zululand Group), and forms the base of the regional aquifer (Fig. 2c).

Overlying the Zululand Group are unconsolidated to partially consolidated sedimentary deposits formed by a succession of marine, alluvial and aeolian processes (Worthington 1978; Meyer and Godfrey 1995; Kelbe et al. 2013; Botha et al. 2013) with varying combinations of sand, silt and clay. The strata have sufficiently different hydraulic properties to form several hydrogeological units that create both unconfined and partially confined (leaky type) aquifers in some locations (Fig. 2c).

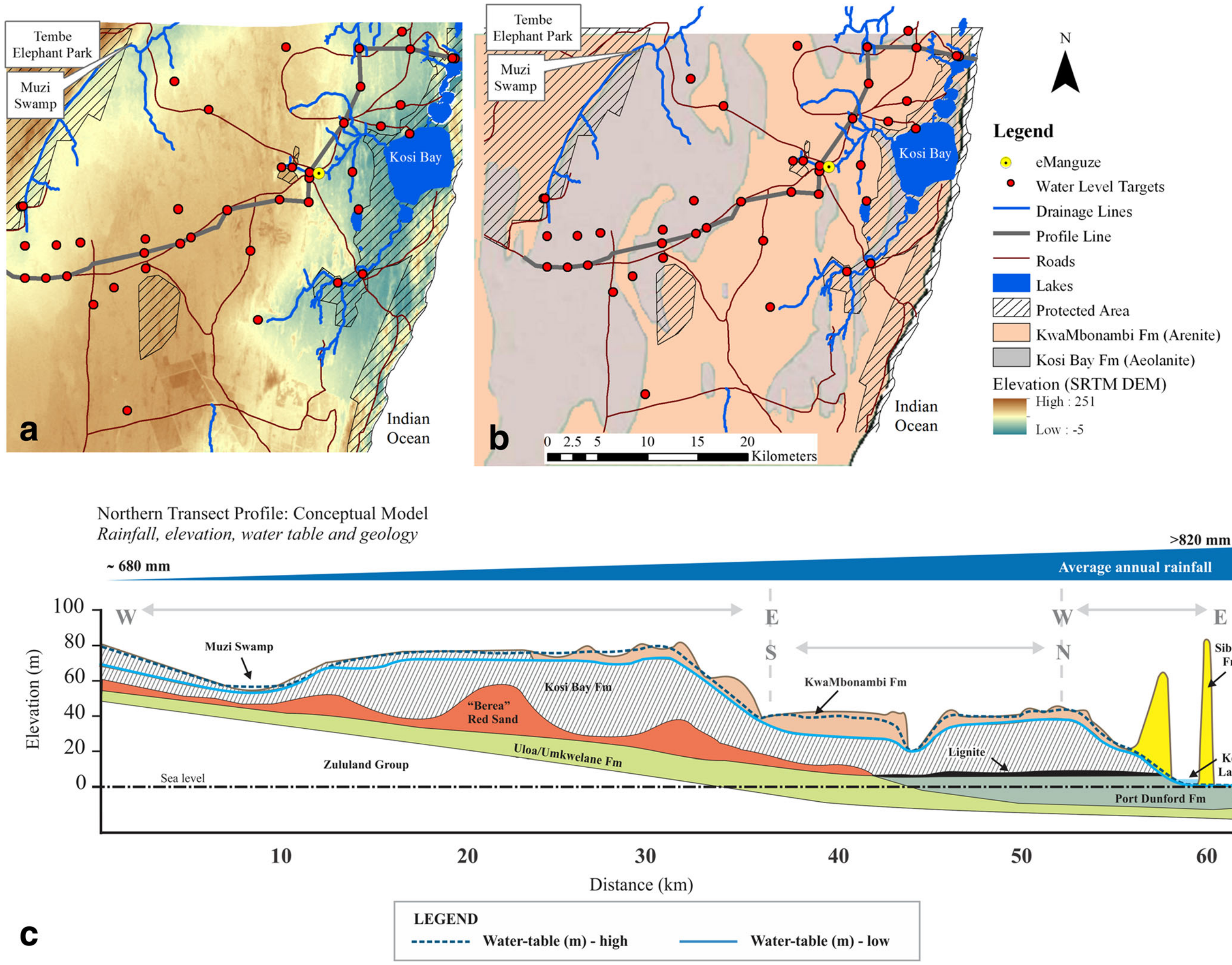

Fig. 2 **a** The regional elevation (*DEM*) derived from SRTM data (CGIAR-CSI 2008; Jarvis et al. 2008). **b** The main geological units that are considered important in regulating the groundwater dynamics in a shallow primary aquifer (Botha and Porat 2007). **c** The inferred schematic geological transect adapted from Kruger and Meyer (1988) (profile line indicated in **a** and **b**) from *Tembe Elephant Reserve* to *Kosi Bay* (Grundling and Grundling 2010)

The lowest part of this primary porosity aquifer (Mio-Pliocene sediments) consists of karst-weathered calcarenites with intercalated mudstone beds (Maud and Botha 2000), often referred to as the Uloa Formation (Fig. 2c). This hydrostratigraphic unit (HSU) (i.e., stratigraphic zones with uniform hydraulic properties) is generally overlain by sedimentary units with finer-grained, less permeable sediments creating a leaky type aquifer (Todd 1980). Along the coastal margin, this overlying unit comprises an extensive layer of middle to late Pleistocene marine, estuarine clay, silt and sand of the Port Durnford Formation. These sediments generally have lower hydraulic conductivities and storativities than the underlying Uloa Formation, forming a partially confined leaky aquifer that is hydraulically connected to the Indian Ocean in places (Kelbe and Germishuyse 2010).

Overlying the extensive middle to late Pleistocene Port Durnford sediments are younger porous and more permeable sandy formations of late Pleistocene to Holocene age. These layers form the Kosi Bay Formation that covers an extensive area from the coast to the western interior (Fig. 2b,c). Separating the Kosi and older formations are interspaced bands of lignite and red sands. The uppermost, youngest Holocene sediments (Sibaya Formation) and the reworked sands of the KwaMbonambi Formation covering a large section of the study area in the eastern regions have relatively high hydraulic conductivity and drain rapidly (DLP 1992). The Sibaya Formation generally occurs in the high frontal dunes above the phreatic zone (DLP 1992) and plays little role in groundwater movement. An overview of the relationship between the geology and hydrogeology of the region has been presented by Lourens (2013).

The spatial distribution of the upper geological units was mapped by Botha and Porat (2007) and plotted in Fig. 2b, along with the available water level monitoring sites (WL targets). There are few borehole logs for the region so it is not possible to delineate the spatial distribution of the vertical profile of the deeper lithological units as portrayed in the conceptual model in Fig. 2c. The water level monitoring points are all shallow wells that measure the water table in the upper lithological units (KwaMbonambi and Kosi Bay Formation) so it is also not possible to calibrate the hydraulic properties of the lower units without additional measurements. Consequently, it has been assumed that the primary aquifer above the low-permeability St Lucia Formation of the Zululand Group forms a single layer in the hydrogeological model. The one layer model formed by the surface topography and the upper surface of the St Lucia Formation (derived from the depth of available boreholes) was used to develop the numerical model using an initial estimate of the hydraulic properties derived for the geological units in Fig. 2b. The final spatial distribution of the hydraulic properties representing the various formations were then derived with a numerical groundwater model using calibration techniques in conjunction with the measured water level at the monitoring sites as described later in the text.

Hydrometeorology

Rainfall

The average annual precipitation in the study area over the last 100 years is 908 mm (CRU 2013), while below-average annual rainfall of 723 mm was measured over the latter 26 years for the period January 1979 to December 2014 (see Fig. 3). The study area lies in the subtropical region with predominantly summer rainfall generated by free convective systems producing high intensity, short duration (sub-daily) storm events (Kelbe 1984; Garstang et al. 1987). Winter rainfall is characterised by larger synoptic systems that produce persistent low-intensity forced convective rainfall that can last for several days over much greater areas. The characteristic size of the convective storms producing significant rainfall in the region is about 10 km^2 and lasts for an average of 20 min (Kelbe 1984). They are significantly smaller than the study area and will recharge only a section of the aquifer during any one event. These convective storms and frontal systems generally produce rainfall events with high rates of precipitation that are less affected by interception (canopy storage) losses. Everson et al. (2014) indicate that under various commercial forestry species, the interception losses in the region are between 10 and 35 % of gross precipitation depending on the canopy storage.

Evaporation and evapotranspiration

The actual evaporation rate varies with potential evaporation rate (atmospheric demand), vegetation type (base line rates and rooting depth) and soil moisture content. The long-term mean and variability in the pan evaporation (assumed to be equivalent to the reference ET or atmospheric demand) in the study area is shown in Fig. 3. The mean annual potential evaporation rate is 1,904 mm (Mucina and Rutherford 2006)

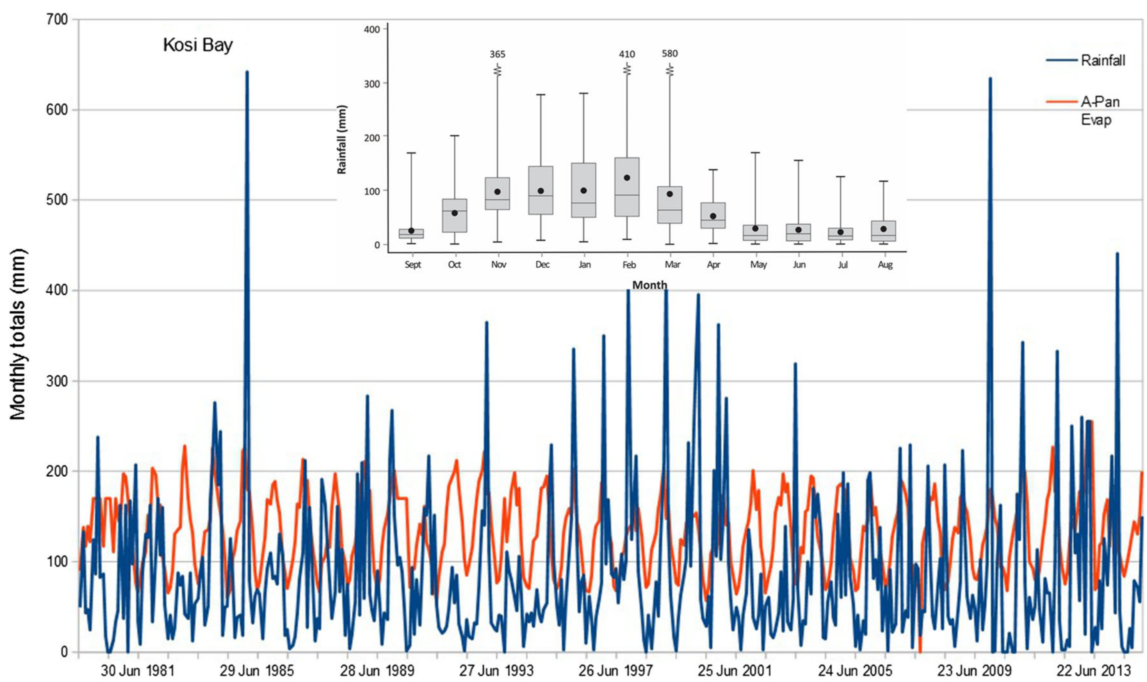

Fig. 3 The monthly rainfall and evaporation series used in the model simulations. *Inset* is the average, minimum and maximum (*box-and-whisker plots*) rainfall over 23 years (Jan 1989–March 2012) arranged according to the hydro-calendar (Sept–Aug) for the study area. The *box* is the range from first to third quartiles in which the rainfall values with the median (*line in box*) and average (*dot in box*); the *whiskers* are the minimum value and maximum values recorded (Grundling et al. 2014)

and the mean annual Class A Pan ET for the past 26 years was 1,596 mm.

Evaporation measurements by Clulow et al. (2012a, 2015) on the Maputaland Coastal Plain have established some baseline rates for different vegetation types that are typical of the study area. Summer and winter evapotranspiration from a permanent peat-dominated wetland was 575 and 325 mm, respectively (900 mm/year) and from a grass-covered dune upland was 303 and 175 mm, respectively (478 mm/year; Clulow et al. 2012a). Their studies indicated that only the Swamp Forest average ET values came close to meeting the atmospheric demand. In others studies, Clulow et al. (2011, 2012b) measured evaporation from commercial pine forests in the riparian zone (i.e., shallow water-table conditions) and determined that the annual evaporation rate of 2.5 mm/day was 75 % of the FAO 56 reference evaporation (Allen et al. 1998) and 94.5 % of the annual rainfall. These values have guided the initial parameterisation of the evaporation model.

Model parameterisation approach

Interception

Interception losses are a function of the rainfall intensity and the land-cover as defined by the leaf area index (LAI). LAI, which is the one-sided leaf area per unit ground area, defines an important structural property of a plant canopy that captures a portion of the incident rainfall for subsequent evaporation, thereby reducing the available rainfall for infiltration and recharge. The proportion of water lost from interception can vary from negligible amounts to almost all the incident rainfall (Calder 1990). The loss is mainly a factor of the canopy storage (estimated from the LAI) and the rate and duration of the incident rainfall. However, leaf morphology and evaporation rate also influence the interception losses. These processes operate at sub-daily rates and require detailed data that are not readily available for this study. Consequently, the interception losses have been estimated as a direct proportion of the LAI for each vegetation type. Land-cover classes (Grundling et al. 2013a) were used to derive the LAI for application in the model (NLC2000 Management Committee 2005). LAI values were derived from MODIS (ESDT: MOD15A2) 8-day Composite NASA MODIS Land Algorithm (Reed 2002). The MOD15 LAI and fraction of photosynthetically active radiation absorbed by vegetation products are available on a daily and 8-day basis provided at 1-km^2 pixel resolution.

The interception storage is incorporated in the model by reducing the gross rainfall incident on each vegetation class by a factor of 10 % of LAI, which induced an interception loss of up to a maximum of 6 mm/day. Since the purpose of the model simulation is to define the seasonal variation in the water-table profile defining the wetland system, it was decided to use monthly rainfall for this study.

Monthly MODIS LAI images from March 2000 to March 2010 were used to derive the LAI. The winter monthly LAI values for April to September were summed for every pixel overlaying a groundwater monitoring point and an average value for every pixel overlaying a groundwater monitoring point was calculated for the winter months. The same was done for the summer monthly LAI values for October to March. Both sets of images (winter and summer) were then used in the creation of zonal statistics from the National Land-Cover 2000 dataset (NLC2000 Management Committee 2005). For each National Land-Cover 2000 polygon, the summer and winter average values were calculated for years 2000 to 2010.

Surface topography and drainage boundaries

There are two main types of drainage boundaries that need to be considered. The vertical processes driven by rainfall and evaporation that occur over the entire surface domain of the region and those fluxes that involve the lateral flow of water, down a hydraulic gradient, to the point of lowest potential energy, which is generally taken as mean sea level (MSL).

Wetlands will form when the water-table elevation is within the rooting zone of the wetland vegetation. Consequently, it is necessary to establish those locations or sites where the depth to the water table is suitable for the development of the wetlands. To establish the depth to the fluctuating water table it is necessary to derive the topographical surface elevation profile in conjunction with the water-table profile. Topographical elevation based on freely available SRTM data (Farr et al. 2007; Hirt et al. 2010) for an area near St Lucia (50 km to the south of Kosi Bay) was found to have vertical elevation errors at pixel resolutions (90 × 90 m) that can exceed 10 m for those areas with tall dense forests; however, the errors were generally within 2 m for areas with short vegetation or bare soil (Carabajal and Harding 2006; Kelbe and Taylor 2011). The existence of commercial Eucalyptus plantations exceeding 10 m in height in the study area causes serious errors in establishing the DEM using the SRTM data. Consequently, an alternate source of elevation data was required. Elevation contours of 5 m were acquired from the National Department of Survey and Mapping for the study area, excluding the section in Mozambique (NGI 2013), where it was necessary to use SRTM data. These data were used to generate the DEM with an estimated vertical accuracy of ±2 m.

The lateral boundaries are formed by groundwater seepage through valley bottoms into streams and rivers that drain to the lakes and sea. These surface drainage systems (i.e., streams and rivers) are characterised by a range of flow rates and residence times that are generally one or more orders of magnitude greater than the groundwater flow rates. In the model domain there are many different types of drainage boundaries that need to be identified and their physical features determined. An error in specifying the elevation of a drainage boundary will be directly transferred to the derivation of the water-table profile. Since the water-table profile is determined by the surrounding drainage boundary features (such as streambed elevation), it is essential to define the external (outer) drainage boundaries that will completely determine the groundwater profile for the area of concern. Internal drainage boundaries will then influence the local variation in the regional groundwater profile. The main focus area of this study was the region between the Kosi Bay lakes and Tembe Elephant Reserve (Fig. 2). The spatio-temporal patterns of water table in this area are strongly influenced by the fluctuation of water levels in the Kosi Bay lakes and the Muzi River/swamp drainage network. While one can assume with some confidence the water level in the Kosi Bay lakes is close to mean sea level, it is not possible to do the same for the swamps. Consequently, both of these boundaries were configured as internal boundaries by extending the model domain to include distant boundaries (Fig. 4).

Modelling approach

Model structure

The derivation of the water-table profile was generated using the MODFLOW 2005 (Harbaugh 2005) model and the suite of accompanying packages that link the shallow unconfined aquifer to the recharge and discharge zones. MODFLOW was configured in two dimensions as a one-layer model with varying depths formed by the surface topography and upper surface of the St Lucia Formation as illustrated in Fig. 2c. This 2D model assumes a no-flow lower boundary and anisotropic hydraulic properties derived during the calibration process. The average depth of the model was 30 m with a minimum thickness of 5 m, while the depth along the frontal dunes could exceed 100 m in places. The initial spatial distribution of the hydrostratigraphic units (zones) was configured according to the geological units shown in Fig. 2b. The initial hydraulic properties of the various hydrostratigraphic units were derived from published values (DLP 1992; Worthington 1978; Været et al. 2009). The internal streams were configured using the SFR1 package of Prudic et al. (2004). The width of the streams increased by 1-m increments from 1 m for first order streams. The streambed slope was derived from the DEM. The streambed roughness values were derived from published

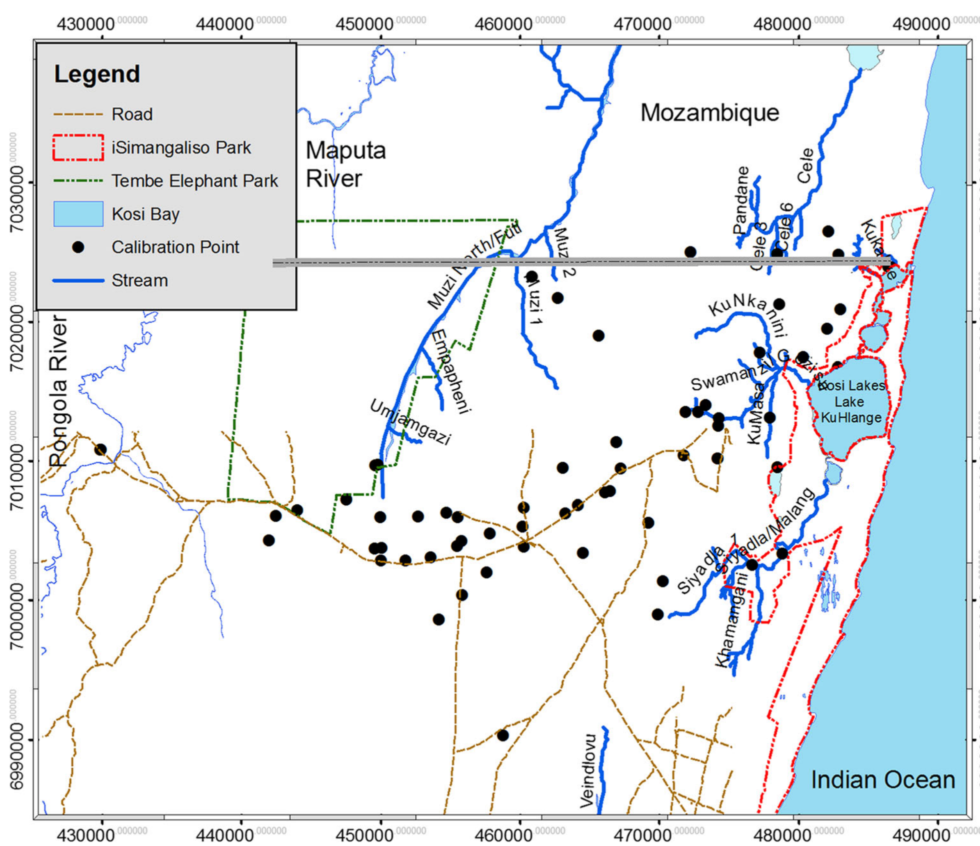

Fig. 4 The model domain and selected drainage boundaries. The northern and southern boundaries were defined as no flow boundaries while the *Pongola and Maputa Rivers* to the west of the Muzi catchment and the *Indian Ocean* in the east formed the other two external boundaries of the model domain. All the other streams were configured as internal boundaries using the SFR package

values for medium to coarse sand channels. The streams were assumed to be free draining so all the hydraulic properties were equated to the aquifer conductivity values. The lakebed elevation was derived from published bathymetric survey and assumed to be free draining so the hydraulic properties were also equated to the surrounding aquifer.

This study presents the simulated results for a transient 10-year period from January 2000 to December 2010 using monthly stress periods and daily time steps. The simulation period includes a wet year, 2000, and several dry years when the average annual rainfall received was below average, e.g., 2002, 2003, 2004 and 2008 (Grundling et al. 2013a). The model was configured on the basis of the conceptual model of the hydrogeology and the vertical (recharge and evaporation) and lateral (rivers and lakes) surface exchange boundaries described in the preceding.

The groundwater model domain covered all the external and internal drainage boundaries shown in Fig. 4. These external boundaries include the Indian Ocean to the east and the Pongola/Maputa rivers to the west of Ndumu. The northern and southern external boundaries are assumed to be zero flux (Neumann type) boundaries, since the primary hydraulic gradients along these external boundaries are perpendicular to the coast. The depth of the aquifer was generally greater than 30 m and the model nodes were 100 m × 100 m. In this study, the groundwater recharge and evaporation were simulated for various land-cover classes (Fig. 1c; Grundling et al. 2013a) using the Unsaturated-Zone Flow Package (Niswonger et al. 2006). The model also incorporated the stream flow package (Prudic et al. 2004) coupled to the lake package (Merritt and Konikow 2000) to simulate the rate of discharge from groundwater storage into surface storage features.

MODFLOW requires detailed description of the hydrogeological features that control the movement of water within the aquifer. These features were initially configured as Hydrostratigraphic Units (HSU) with homogeneous hydraulic properties (hydraulic conductivity and storativity) but spatial heterogeneity was introduced using the PEST inverse calibration modelling techniques (Doherty et al. 2010).

Infiltration across the model domain is derived from the effective rainfall (after subtracting interception losses based on the vegetation type (LAI) as described previously). The recharge to groundwater storage is derived from the infiltrated water after evaporation/transpiration losses from the unsaturated zone have been satisfied using the Unsaturated-Zone Flow (UZF) Package (Niswonger et al. 2006). In this model, the infiltration rate is limited by the vertical hydraulic conductivity (Kv), soil moisture characteristics and the Brooks-Corey relationship between unsaturated hydraulic conductivity and water content of the upper soil profile (Niswonger et al. 2006). Excess soil moisture (above the field capacity) is routed to surface streams as direct runoff by the UZF package. The soil moisture in the unsaturated rooting zone is further depleted by evapotranspiration. The UZF routine uses a kinematic wave approximation to Richards' equation to simulate vertical unsaturated flow of the wetting front (Niswonger et al. 2006) in the unsaturated zone. The UZF routine extracts the evaporation component at the potential rate derived from the atmospheric demand (pan evaporation) and a factor for each vegetation class. The evaporation is extracted from the unsaturated zone above the extinction depth (rooting depth) for each vegetation class (Fig. 1c). If the evaporative losses in the unsaturated zone are less than the atmospheric demand then evaporation continues from the saturated zone whenever groundwater is within the rooting zone (extinction depth) derived from literature studies (Canadell et al. 1996; Benyon et al. 2006; Schenk and Jackson 2002). All remaining soil water after satisfying the evaporative demand percolates into the saturated zone.

Advanced studies of certain boundaries such as lakes require much more complex conceptual and numerical models that have to account for the mass balance involving all sources and sinks. The lakes have been incorporated into the model using the LAK3 package of Merritt and Konikow (2000). This model requires, amongst other fluxes, the stream flow into and out of the lakes, which was simulated using the SFR1 package of Prudic et al. (2004).

Numerical modelling

The initial specification of model parameters was derived mostly from other studies of the Maputaland coastal environment (e.g., Kelbe 2009; Kelbe and Germishuyse 2010; Været et al. 2009; Kelbe et al. 2013).

The streambed elevation forming the drainage boundary was derived from the DEM and then lowered by 0.5 m to mimic the shallow stream channels forming the drainage lines. The lake bathymetry was obtained from Wright et al. (2000). The streambed roughness (Manning coefficient) and width was based on site visits to several access points near road crossings. The Manning coefficient was taken from the website at The Engineering ToolBox (2015). The measured water level in the central lake was obtained from the South African Department of Water and Sanitation (DWS).

Sensitivity analysis was manually performed on the recharge and evaporation parameters used in the UZF package and selected parameters in the STR1 and LAK3 packages using the steady-state model. Interception losses (effective rainfall) and the saturated and unsaturated hydraulic conductivity parameters were selected for the model calibration. The aquifer, lake and stream parameters were configured and calibrated manually by adjusting their hydraulic properties to achieve agreement between the simulated and measured water level data in wells, wetlands, streams and lake and the results analysed using the Groundwater Vistas Interface. The calibration process systematically changed the model parameters to achieve the best agreement between the measured and predicted values of model variables. For the groundwater storage, the calibration was based on measurements for the water levels in the monitoring wells shown in Fig. 4. The hydraulic properties were derived using the Model-Independent Parameter Estimation (PEST) techniques developed by Doherty et al. (2010). However, before applying PEST, an attempt to calibrate the recharge rate to the aquifer was conducted by systematically adjusting the recharge rates (interception and infiltration) to achieve an acceptable balance with the discharge rates and change in storage of the groundwater (as measured by the water-table elevation and the water levels in the lakes). The discharge from the aquifer is through the land surface (evaporation) and seepage along the drainage lines forming the streams and lake shorelines. These discharge rates are calculated by the model and should be validated against measured flow rates where possible. However, no runoff measurements have been recorded for any of the streams in the study area, so it was not possible to calibrate the recharge using direct runoff measurements. Nevertheless, the lake model requires the stream flow into and out of the lake to balance all the other fluxes with the change in storage. There is no known abstraction from these lakes so it is assumed that the change in storage is due to the natural fluxes comprising rainfall, evaporation, runoff and groundwater seepage. The rainfall and evaporation rates were applied directly to the lake surface area. Lake water-level measurements (change in volume) have been recorded by the DWS at sub-hourly rates for the simulation period. It is assumed that good agreement between the simulated and measured lake water storage (±0.25 m) signifies reasonable rates of inflow and outflow from the stream and groundwater. These discharge rates from the system must be in balance with the recharge rate to the regional aquifer if the change in water table (groundwater storage) is close to the measured water table.

If the simulated water levels that are in agreement with measured lake level signify acceptable discharge rates from the drainage boundaries, then it is assumed that the recharge and evaporation rates are an acceptable estimate for this study. The recharge rates were then applied over the simulation period to derive a spatial distribution of the hydraulic properties using the PEST calibration technique. A regular 2,000-m separation of pilot point over the extent of the model domain with supplementary pilot points around head targets were used in the SDV approach in PEST with regularisation (Doherty et al. 2010) to derive a spatial estimate of the hydraulic conductivity. The derived hydraulic conductivities were generally lower (<20 m/day) in the eastern part of the study domain, with the exception of some paleodune ridges, and higher in the Muzi Valley and toward the western part of the study area.

The high correlation ($r^2 = 0.99$) between the simulated and measured groundwater and lake-water levels (Fig. 5) is considered adequate to provide the best estimates of the water balance of the system and adequately represent the groundwater storage in the aquifer for the evaluation of the spatial and temporal changes in the depth to the water table for this study. While an attempt was made to validate the recharge rate, the derived estimates of the recharge and the hydraulic properties of the underlying aquifer are not necessarily unique and represent only one configuration of the driving variable and aquifer parameters. While other combinations of the recharge and hydraulic properties will give similar agreement between the measured and simulated water-table profile in the aquifer, the only necessary criterion for using the model predictions in this study is its ability to simulate the water-table profile. The model is not used for simulating other conditions that would rely on a unique set of parameters. Consequently, it is argued that the model configuration is sufficient to derive the best estimate of the depth to the water table necessary to describe the geomorphic setting for the formation of wetlands. The accuracy of the evaluation also rests on the reliability of the topographical surface (DEM).

Results

This study has adopted the concept that the depth to the water table is the main criterion for the development and sustainability of wetlands of various forms. Consequently, the model evaluation is based on an assessment of the correspondence between the spatial distribution of the classified wetland types derived by Grundling et al. (2013a) and the spatial distribution of the proposed minimum water-table depth that will support the main wetland types in the coastal aquifer.

Model results

The main drivers in the prediction of groundwater storage (water-table profile) are the recharge and lateral discharge processes. The discharge to the streams has been evaluated as part of the lake water balance by adjusting the hydraulic parameters to achieve correspondence between the simulated and measured lake storage (water levels). The model overpredicts the lake level by about 10 cm (0.1 m). The water balance for the entire model domain using the steady-state model (average conditions) indicates that the actual evaporation loss is 78 % of the recharge rate with the remaining 22 % lost through drainage systems (rivers, lakes and ocean).

Delineation of wetland area

The mean depth to the water table for the simulation period from January 2000 to December 2010 is shown in Fig. 6 for locations where the water table was no deeper than 2 m below ground surface, where permanent or semi-permanent wetlands

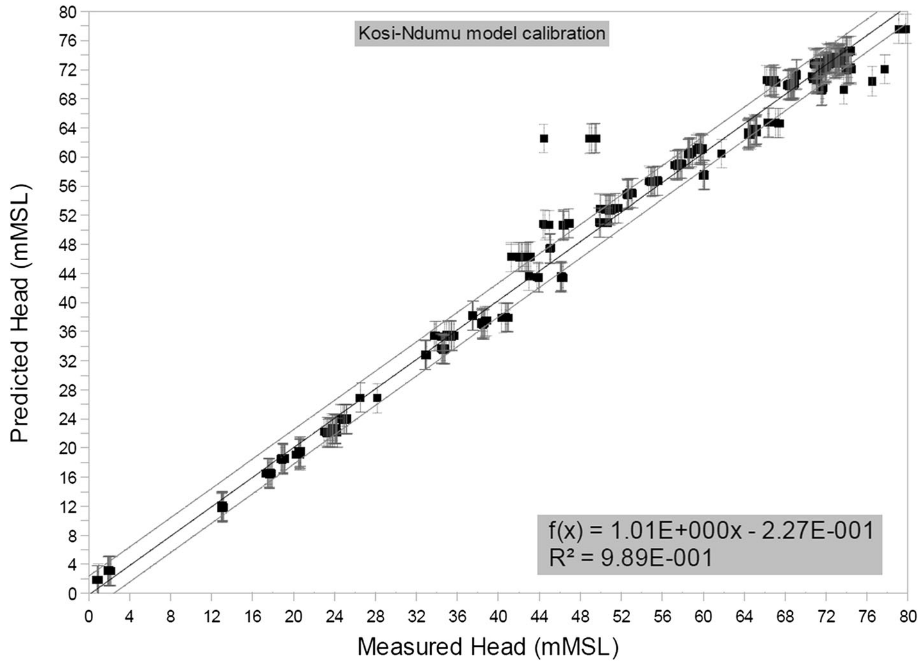

Fig. 5 Model generated head predictions (2000–2012) plotted against measured heads. The *error bars* represent a 2 m range for the measured values. Least squares fit gives an $r^2 = 0.99$ and a slope of 1.01

were expected to form (Grundling et al. 2013a). The model shows specific regions of the study area along drainage boundaries with shallow water table that are likely to be supportive of wetland vegetation. The distribution of these simulated wet areas compares favourably in many areas to the wetlands classified by Grundling et al. (2013a). Generally the 2-m contours of the depth to the water table show the expected close correspondence with the wetlands in the low-lying river courses in the Muzi system along the Tembe Elephant Park boundary and to the south of Kosi Bay (Lake KuHlange; compare Fig. 6 with Fig. 1c).

As expected, the water-table fluctuations are very small (standard deviation <0.1 m) near the streams and lakes (Fig. 7) due to the static nature of these discharge boundaries. However, there are large fluctuations of >1 m standard deviation in the aquifer between these drainage boundaries that imply changing water levels of >2 m during the simulation period. Since the type and form of the wetlands are controlled

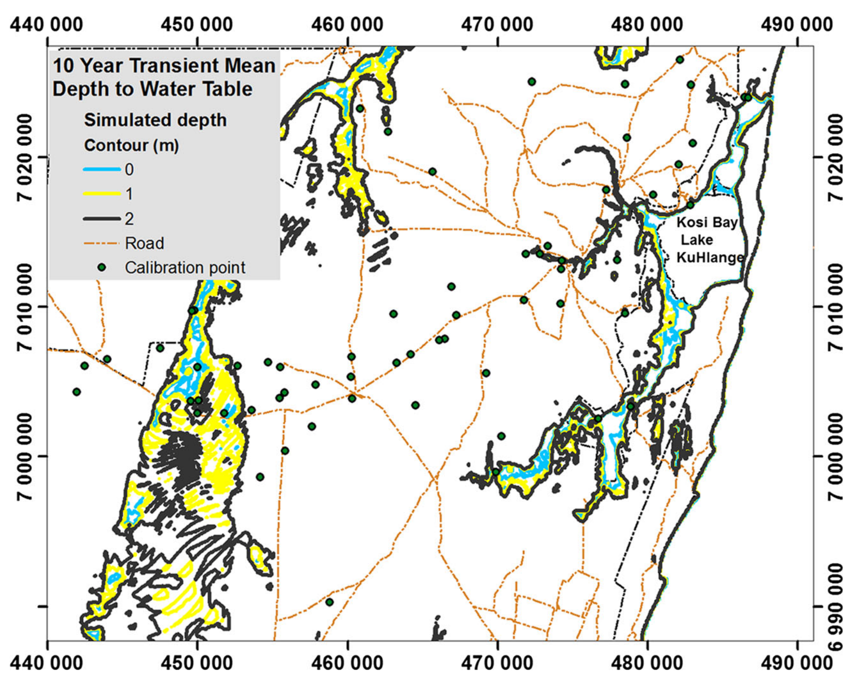

Fig. 6 The location of the areas where the simulated depth to the water table was <2 m under average conditions during the 10-year simulation period. Contours units are metres below ground surface

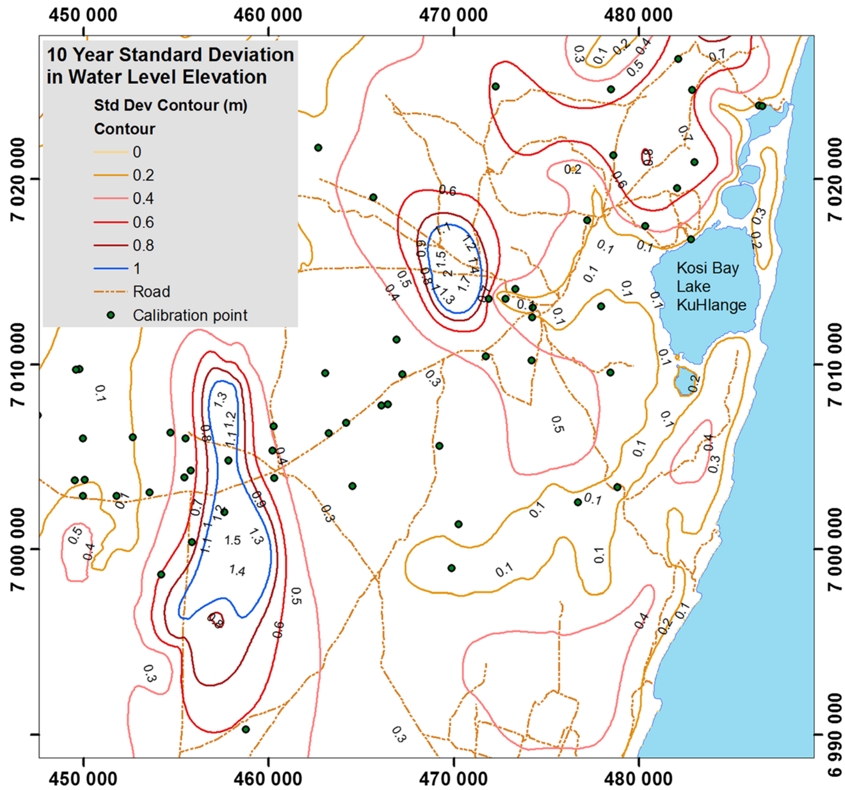

Fig. 7 The standard deviation of the simulated depth to the water table (metres below ground surface)

by the fluctuation in the water table, these areas are likely to have more transient types of wetlands as is shown in Fig. 1c. The zones of high variability are associated with those areas for which low hydraulic conductivity values were derived in the model calibration. Although many of these zones are close to monitoring points, the localised zone of high fluctuation directly west of Lake KuHlange (see Fig. 7) may be an artefact of the calibration process where there are no monitoring points.

The minimum and maximum water-table profile for the simulation period (2000–2010) was extracted from the simulation series and used to compare the wetland distribution under dry and wet conditions (Figs. 8 and 9). Figure 8 shows the predicted areas where the water table is shallower than 2 m below ground surface for this dry period. This minimum water-table profile is assumed to represent the spatial distribution of the limit of suitable hydrological conditions for wetlands under prolonged dry conditions. During these periods, temporary wetlands become dry and some permanent wetlands become reduced in apparent size (Grundling et al. 2013a), so were not observed in the remotely sensed image. The model indicates greatly reduced areas favourable for wetlands between wet and dry periods along the Muzi Valley and in the vicinity of the Kosi Bay lakes during wet and dry periods. In the upland plateau between the Muzi and Kosi Bay lake drainage systems the model shows very few areas where the regional water table is higher than 2 m below the surface, implying the area is generally unfavourable for permanent wetlands linked directly to the regional groundwater.

The simulated depth to the water table for the wet period (Fig. 9) shows the area with shallow groundwater (<2 m depth) has greatly expanded and covers large areas of the upland between the Muzi Valley and Kosi Bay lake systems. These represent areas more favourable for wetlands to occur in wet periods, thus are likely locations for the development of temporary wetlands.

The simulated average 2-m contour of the depth to the water table was overlaid on the classified wetland types (Grundling et al. 2013a) for 2008 (Fig. 8) and 2000 (Fig. 9) and used to identify most of the wetlands within the river valleys. However, the 2 m contour did not extend to the extensive area between Tembe Elephant Park and the Kosi Bay catchment where a predominance of temporary wetlands occur (central part of Fig. 9). The selection of a 3-m contour may have included some of these regions. Alternatively, differences in the clay fraction of surface soil horizons can affect wetland occurrence (e.g., in temporary upland wetland systems) but the model was unable to simulate these with the available soil data.

Further examination of the wetlands located in the study area shows clear relation to the topographic features and clay content. The clay occurrence map (Van den Berg and Weepener 2009; Van den Berg et al. 2009) is shown in Fig. 10a and indicates that the weathered clay-enriched soil

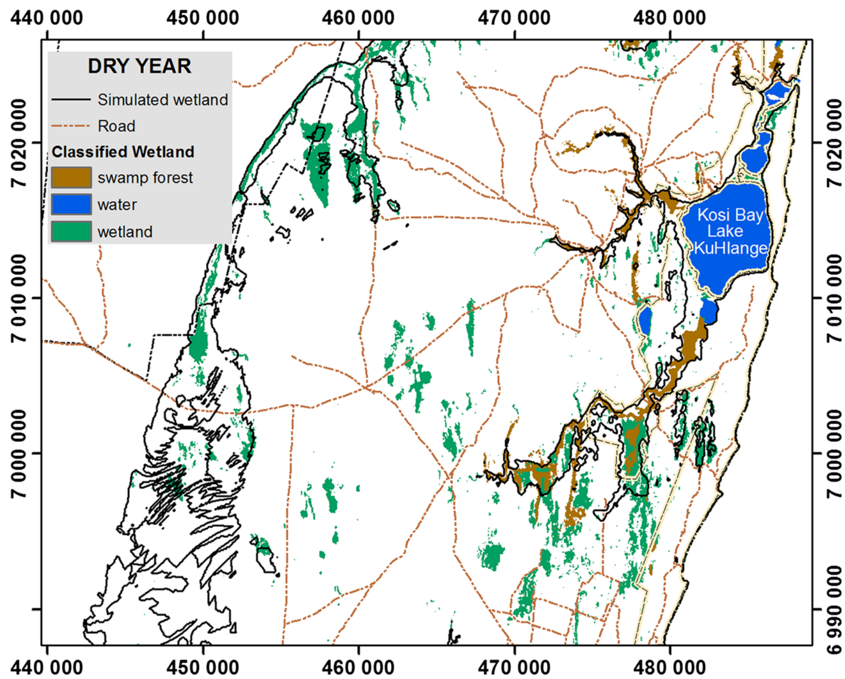

Fig. 8 The classified wetland distribution during a dry period (2008). Superimposed on these images is the predicted 2 m depth to the water-table contour for corresponding dry conditions (i.e., the maximum depth to the regional water table during the 10-year simulation period from 2000 to 2010)

found in soil profiles corresponds well with the wetlands mapped in Fig. 1b. Grundling et al. (2014) compared the clay occurrence with wetland distribution and indicated that ~49 % of permanent wetlands in the study area are associated with areas with soil having >16 % clay content. In contrast, ~63 % of wetlands occur on soil with <5 % clay, and correspond with the distribution of temporary wetlands (Grundling et al. 2014).

It is concluded that most temporary wetlands (those that fall within the 2-m water-table depth during the wet period; Fig. 9) are linked to the regional water table, generally being associated with low % clay occurrence. At some temporary wetlands, notably those that occur in the central upland plateau outside the 2 m water-table depth contour, it is likely that lower hydraulic conductivity caused by higher clay content, buried ferricrete or paleo-peat layers contribute to a prolonged hydroperiod (Grundling et al. 2014). In wet years with prolonged wet periods, these wetlands could also be connected to the regional water table. As the regional water-table

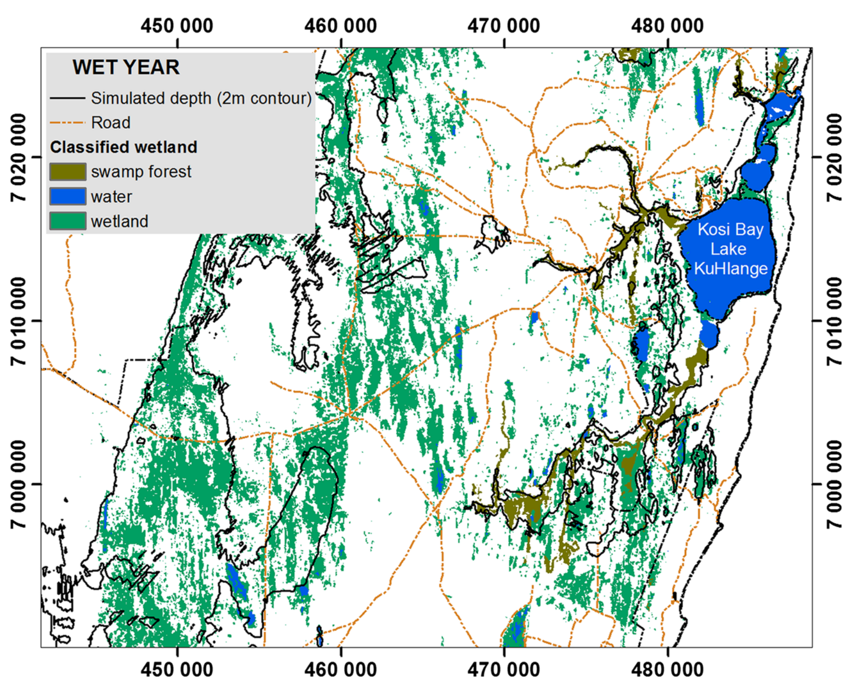

Fig. 9 The classified wetland distribution during a wet period (2000). Superimposed on these images is the predicted 2 m depth to the water-table contour for corresponding wet conditions (i.e., the minimum depth to the regional water table during the 10-year simulation period from 2000 to 2010)

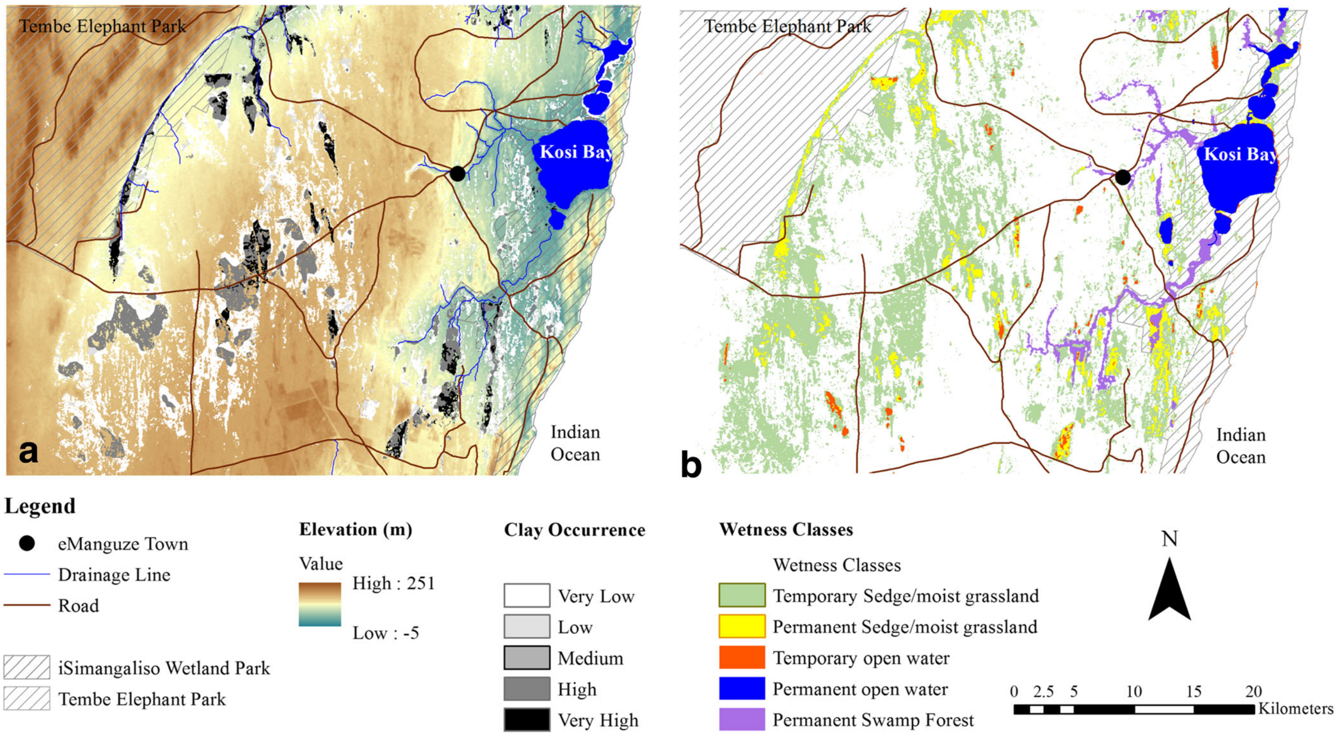

Fig. 10 The **a** clay occurrence and **b** wetland distribution for the area between the Muzi and Kosi Bay drainage systems (Grundling 2014)

subsides, perched water apparently persists on lenses of fine-grained sediments in the soil profile. More hydrological detail is needed to substantiate this.

Discussion and conclusions

This study examines the use of a numerical groundwater model to predict the water-table conditions in the north-eastern Maputaland Coastal Plain, under wet and dry conditions, to improve our understanding of the relation between water table and the spatial distribution of wetlands and their temporal behaviour. The simulated water-table profile is considered alongside previously identified temporary and permanent wetland and open water areas to provide a more comprehensive understanding of the groundwater-dependent ecosystem.

Under dry climatic conditions (2008), areas with water table shallower than 2 m below ground surface showed a strong correspondence with permanent wetlands (Fig. 8) previously identified (Grundling et al. 2013a). These areas were strongly associated with Muzi River and Kosi Bay lake fluvial systems, where a high and steady regional water table dominates (Grundling et al. 2014). These wetlands typically contain peat, which indicates that they remain in a state of sustained saturation (Grundling et al. 2013b, 2014), and low water-table variability (Fig. 7). The model was also used to predict water-table depth less than or equal to 2 m below the surface during a wet period (2000) (Fig. 9). The area encompassed by the 2 m water-table depth contour was considerably larger than that simulated for the dry period, extending into the upland zone that primarily supports temporary wetlands (Grundling et al. 2013a). However, the zone of temporary wetlands extended well beyond the 2 m water-table depth contour, in an area where the surficial deposits are from the lower permeability sediments of the Kosi Formation, and an area of gently undulating landforms with extensive flat-bottomed features on the upland areas (Grundling et al. 2014), many of which have soils with a high clay content (Fig. 10a). The deeper regional water table associated with these areas suggests that wetland processes rely on transient perched conditions that occur during wet seasons and especially during wet years (Grundling et al. 2013a). To capture the zone of temporary, perched wetlands, a more sophisticated saturated-unsaturated modelling approach is required, along with representation of the layered heterogeneity of soils associated with the different formations and soil types that occur. Nevertheless, the groundwater simulations done here highlight the temporary wetlands that are most likely disconnected from the regional water table, and thus more susceptible to climatic, and perhaps local anthropogenic stressors such as plantations. The presence of these temporary wetlands above the regional water table in the upland plateau (Fig. 2c) illustrates their importance for depression-focused recharge (Berthold et al. 2004; Derby and Knighton 2001), which in part provides water to wetlands at lower elevations (Winter and Rosenberry 1995), in this case the peat-dominated Muzi wetlands and Kosi Bay swamp forests.

Hydrological models contain inherent uncertainties and weaknesses. Schultz (2013) considers that the projections of hydrological models, as numerical abstractions of the complex systems they seek to represent, suffer from epistemic uncertainty due to approximation errors in the model, incomplete knowledge of the system, and in more extreme cases, flawed underlying theories. Faulty or inaccurate data (e.g., biased water-table data or inaccuracies in the DEM) used for development, calibration or validation can also be problematic. However, where the model is used as a simple tool for defining the water-table profile of an area where sufficient hydrological and geological information is incorporated, the level of uncertainty can be acceptably low. While the model in this study has been used solely to predict the water-table profile, it is acknowledged that the model parameter set is not unique, and there is a high likelihood that other parameter sets will provide an equally good representation of the water-table profile. Therefore, no attempt has been made to validate the model hydrodynamics, which would require a priori knowledge of transient processes. Consequently, the main concern with the model is the accuracy of the water-table prediction compared to measured values (Fig. 5) and the suitability of extrapolation to areas with little or no monitoring. Here, Kriging functions (PEST) were used to extrapolate hydraulic properties so that model predictions of water table could be made in areas where no measurements are available. The relative vertical accuracy of the elevation data is up to 1 m, which adds additional uncertainty in the evaluation of shallow water table depths below the ground surface.

Within the constraints imposed by the inherent inaccuracies and availability of field data in this study, it is believed that the numerical methods used to estimate the water table profile in this shallow unconfined aquifer provided acceptable predictions for the purpose of delineating zones where the permanence of wetlands can be explained. However, the use of remote sensing methods and applications to enhance the derivation of model parameter sets could greatly improve the model predictability in this type of study.

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