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A REVIEW OF PROCEDURES FOR WATER BALANCE MODELLING

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Water balance is another name for the principle of mass conservation in which changes of total water volume, inflow (precipitation, snow melt) and outflow (evaporation, transpiration, surface and subsurface runoff) on a given area are balanced. The study of the water balance with a previous knowledge of climatic and physical basin characteristics offers information about current and future water quantities, and added insight into the complex process of basin runoff. This paper gives a review of methods for defining water balance and its main components, a review of numerical models for water balance calculation and examples of model applications.

INTRODUCTION

The movement of water and moisture throughout the continuum of the ground, vegetation and atmosphere is important for the human, plant and animal world. Knowledge of total water inflow and outflow from a catchment gives insight into present and future available water storage and may help in defining variants for water management strategy. Satellite and aerophotogrametric measuring techniques have enabled greater insight into the physical and biophysical processes which control water balance. However, the estimation of water flux between water balance components is still an interesting and demanding hydrological challenge and the effects of land use change on water balance and estimating runoff in unstudied watersheds are continuous themes of scientific and expert ecohydrological studies (Zhang et al., 2008; Todini, 2007; Alemaw and Chaoka, 2003).

There are various research and applied problems where the calculation of water balance is used: for the estimate of a regional water balance, for the assessment of the impact of human activity and climatic variations on basin runoff, in planning and allocation of fresh water resources, in engineering applications such as bridge management systems, etc. An understanding of water balance in relation to climatic and morphological basin features gives us insight into complex processes which are conducted regarding different spatial and temporal relations (Zhang et al., 2008). The predictions of more frequent and longer drought periods and of greater intensity of floods clearly define the need for a more detailed knowledge of existing and of future watershed conditions. For such knowledge the calculation of water balance is essential and may also provide reliable information in defining strategies for climate change mitigation measures on the watershed.

Different mathematical hydrological models have been developed which vary in relation to; the general modelling approach and methods of calculating different water balance components, the complexity of model structure and the applicability to different areal and temporal domains, etc. The complex interconnected links between individual components of the hydrological cycle require complex mathematical approaches, on one hand, and hydrometeorological and geophysical datasets on the other hand. Since the required input data are often unavailable or unreliable, it is useful to find a balance between the detailed parameterisation of processes and the quality of input datasets.

This paper gives a review of methods for defining water balance and its main components, a review of numerical models for water balance calculation and examples of model applications.

APPROACHES TO THE MODELLING OF HYDROLOGICAL PROCESSES

The history of hydrological modelling of runoff extends from the rational method (1850) up to contemporary mathematical water balance models (Todini, 2007). Hydrological models may be classified according to different criteria. Based on a description of hydrological processes there are conceptual and physically based models respectively, and based on a spatial description of the runoff process there are distributed and lumped models.

The first concept of the distributed, physically based model was developed by Freeze and Harlan in 1969. Progress in computational and GIS technology enabled rapid development of detailed spatially based hydrological models which have become significant in contemporary hydrological practice. To date, different models have been developed, for example Thales Model (1992), MIKE SHE (1995), TOPMODEL (1995), SHETRAN (2000), tRIBS (2003) etc (Todini, 2007; Refsgaard, 2007). The use of distributed physically based hydrological models with reasonable simplifications has proved suitable for resolving complex hydrological problems on a larger scale (Alemaw and Chaoka, 2003). Examples of lumped conceptual models are the Stanford Watershed Model (1966), the Sacramento Model (1973), the Xinanjiang Model (1977), the ARNO Model (1996) etc. In practice the lumped conceptual models and distributed physical models are most commonly used.

A specific issue is the application of distributed models on large basins. A large number of grid points require a long computational time, and the number of calculated parameters may be significantly greater in the distributed than in lumped models (Refsgaard, 2007). The solution is either to reduce the number of grid points or to reduce the parameterisation of processes. In such a way the local spatial differences may not be correctly modelled and may decrease the model results quality, specifically the description of surface flow and infiltration connections (Beven, 2001). Another issue may be due to the unavailability of required input datasets. In such circumstances one requires the application of black box models (Beven, 1996) and/or estimations of inputs from climatic functions and from basin morphological features (Zhang et al., 2008).

APPROACHES TO THE DEFINITION OF WATER BALANCE

The general water balance equation is the application of the continuity equation to a basin and can be written as:

$$P=\left(\frac{dS_I}{dt}+E_I\right)+\left(\frac{dS_O}{dt}+E_O+Q_O\right)+\left(\frac{dS_{PP}}{dt}+ET+E_{PP}+Q_{PP}\right)+\left(\frac{dS_P}{dt}+Q_P\right) \tag{1}$$

where P is precipitation, and the following processes are separated in the rounded brackets: on vegetation (S_I – interception storage, E_I – evaporation of interception), surface processes (S_0 – surface water storage, E_0 – surface evaporation, Q_0 – surface runoff), subsurface processes (S_{PP} – subsurface storage, ET – evapotranspiration, E_{PP} – soil evaporation, Q_{PP} – subsurface runoff) and underground processes (S_P – underground storage, Q_P – groundwater flow).

Approaches to the definition of water balance have different model complexity and different numbers of water balance components. Zhang defines water balance by using only the basic components of, total storage, precipitation, evapotranspiration and total runoff (Zhang et al., 2008). Xu adds the soil moisture change component to Zhang's equation (Xu and Chen, 2005) and Alemaw includes infiltration as an additional parameter (Alemaw and Chaoka, 2003). Evapotranspiration may also be treated in more detail as Chen proposes when he separates upper and lower evaporation of interception and upper and lower soil transpiration (Chen et al., 2005).

In the case of an analysis for irrigation needs, irrigation can be added to the water balance equation (Sanchez et al., 2010; Portughese et al., 2005). Also, when modelling subsurface runoff the soil component can be divided into an unsaturated and a saturated zone (Kerkides et al., 1996).

Due care should be given to the selection of the time step for calculations. It is interesting to note that the calculation of the monthly water balance with a monthly time step showed equal or even greater reliability than calculations with a daily time step. For example, on the basis of model results from over 300 Australian watersheds, Wang supports simulations with monthly calculated increments for the calculation of water balance where the primary interest is monthly, seasonal and annual runoff volume (Wang et al., 2011).

ESTIMATION OF BASIC WATER BALANCE COMPONENTS

Potential evapotranspiration

Kerkides and Xu cite Thornthwaite's method as being widely used for the calculation of potential evapotranspiration by using average monthly air temperature (Kerkides et al., 1996, Xu and Chen, 2005). Such methods were proposed by Blaney and Criddle (1950), Thornthwaite (1948) and Hamon (1961) (Horvat, 2012):

$$ET_P = 16\left(\frac{l_i}{12}\right)\left(\frac{N}{30}\right)\left(10\,T_\alpha / I\right)^\alpha \tag{2}$$

where l_i is actual length of day; N is number of days in month; T_α is average monthly air temperature (°C); α is empirical coefficient; l is heat index.

Other methods use precipitation (e.g. Turc method, 1954), solar radiation (e.g. Jensen and Haise method, 1963), air humidity, wind speed and characteristic vegetation cover for the estimation of evapotranspiration (Horvat, 2012). Douglas suggests using Turc's method for areas where relative air humidity is greater than 50% and states (Douglas et al., 2009):

$$\lambda \cdot \rho_w \cdot ET_P = 0.369\,\frac{T_{avg}}{T_{avg}+15}\left(2.06\,R_s + 50\right) \tag{3}$$

where ET_P is potential evapotranspiration (mm/day); λ is latent heat of evaporation (MJ/kg); ρ_w is water density (kg/m^3); R_s is daily solar radiation (W/m^2); T_{avg} is average daily air temperature (°C).

In addition to air temperature and shortwave radiation, Penman (1948) introduced relative air humidity and wind speed as input data. Evapotranspiration from bare and humid ground or from grass covered ground was expressed as a fraction of evapotranspiration from open water surfaces. Kerkides and Polhamus cite the example of Penman's equation for the calculation of potential evapotranspiration (Kerkides et al., 1996; Polhamus et al., 2013):

$$\lambda \cdot ET_P = \frac{\Delta\left(R_n - G\right) + \rho\, c_p\left[e_s\left(T_a\right) - e_a\right]/r_a}{\Delta + \gamma} \tag{4}$$

where λ is latent heat of evaporation (MJ/kg); Δ is pressure gradient of saturated water vapour (Pa/K); R_n is net radiation (W/m^2); G is ground heat flux (W/m^2); ρ is air density (kg/m^3); c_p is specific heat capacity of air (J/kgK); e_s is the saturation vapour pressure (Pa); e_a is actual vapour pressure (Pa); r_a is aerodynamic resistance to transfer of water vapour from the surface to ambient air (s/m); γ is psychometric constant (kPa/K).

Numerous scientists have continued to develop Penman's method to customise it to surfaces covered with vegetation with the introduction of new parameters (aerodynamic resistance, surface resistance), but most commonly a modification of Monteith's (1965) is used. The Penman-Monteith equation, described in (Alexandris et al., 2006; Biftu and Gan, 2000; Chen et al., 2005; Douglas et al., 2009; Polhamus et al., 2013) is:

$$ET_P = \frac{0.408(R_n - G) + \gamma\left(\frac{900}{T} + 273.16\right)U_2 \cdot VPD}{\Delta + \gamma(1 + 0.34\,U_2)} \tag{5}$$

where R_n is net radiation (MJ/m^{2}h); G is ground heat flux (MJ/m^{2}h); γ is psychometric constant (kPa/°C); T is average hourly air temperature (°C); U_2 is wind speed at a height of 2 metres (m/s); VPD is water vapour deficit (kPa); Δ is pressure gradient of saturated water vapour (kPa/°C).

The European Union research body have compared the Penman-Monteith with 9 other methods of evapotranspiration (Choisnel et al., 1992), and have suggested the Hargreaves-Samani formula (1985) as the most suitable. Xu and Alexandris cite Hargreaves' potential evapotranspiration calculation method, which states (Xu and Singh 2005; Alexandris et al., 2006):

$$ET_P = a R_a T D^{\frac{1}{2}} (T_a + 17.8) \tag{6}$$

where a is constant (a=0.0023); R_a is solar radiation (mm/day); TD is maximum and minimum daily temperature difference (°C); T_a is average daily temperature (°C).

One of the Penman's equation modifications is Makkink's model (1957). Xu cites the Makkink method which was later perfected by Hansen (1984) and now states (Xu and Chen 2005):

$$ET_P = 0.7 \frac{\Delta}{\Delta + \gamma} \frac{R_S}{\lambda} \tag{7}$$

where Δ is saturated water vapour pressure (mbar/°C); R_S is total solar radiation (cal/cm^{2}day); γ is psychometric constant (mbar/°C); λ is latent heat of evaporation (cal/g).

Priestley and Taylor (1972) suggest the calculation of evapotranspiration based on average air temperature and solar net radiation, in the following way (Xu and Chen 2005; Zhang et al., 2009; Douglas et al., 2009):

$$ET_P = \alpha \frac{\Delta}{\Delta + \gamma} \frac{R_N}{\lambda} \tag{8}$$

where α is coefficient; Δ is gradient of saturated water vapour pressure (mbar/°C); R_N is net radiation (cal/cm^{2}day); γ is psychometric constant (mbar/°C); λ is latent heat of evaporation (cal/g).

Douglas compared measured daily evapotranspiration with the results from Turc, Priestley-Taylor and Penman-Monteith methods (Douglas et al., 2009). Turc and Priestley-Taylor methods were better in a yearly estimate, while the Priestley-Taylor method proved more satisfactory on daily estimates.

Xu compared seven models of calculating evapotranspiration of which four were for potential evapotranspiration (Thornthwaite, Hargreaves, Makkink and Priestley-Taylor methods) and their influence in determining water balance (Xu and Chen 2005). The results demonstrated that for calculating actual evapotranspiration the Makkink model gave better results than other models. Further, for the calculation of soil moisture, four of the seven models, from which three are used for calculating potential evapotranspiration (Thornthwaite, Makkink and Priestley-Taylor) gave equally good results. Xu concludes that the components of water balance (actual evapotranspiration, groundwater recharger and soil moisture) may be predicted with satisfactory accuracy with the help of the Makkink model.

Actual evapotranspiration

Of all energy driven flows, actual evapotranspiration is the most difficult to measure. For direct measurement it is possible to use evaporation from the water surface, although such measurements are not appropriate for the influence of vegetation on moisture loss. The most frequent method for assessing actual evapotranspiration is the application of analytical and empirical equations based on field measurements. These are developed using correlated measured evapotranspiration and climatological parameters which act directly or indirectly on evapotranspiration (Horvat, 2012).

Zhang cites the method for calculating average yearly evapotranspiration which Fu (1981) proposed (Zhang et al., 2008):

$$\frac{ET_A}{P}=1+\frac{ET_P}{P}-\left[1+\left(\frac{ET_P}{P}\right)^w\right]^{1/w} \tag{9}$$

where ET_A is actual evapotranspiration (mm/day); P is precipitation (mm); ET_P is potential evapotranspiration (mm/day); w is model parameter range $(1,\infty)$.

Portughese calculates indirect actual evapotranspiration based on potential evapotranspiration, for each month separately as (Portughese et al., 2005):

$$ET_A=f_i \cdot K_c \cdot ET_P \tag{10}$$

where ET_P *i*s potential evapotranspiration (l); K_c is monthly crop coefficient (-); f_i is function of water extrapolation from the ground (-).

Methods for direct calculation of actual evapotranspiration are somewhat more complex. Xu cites the so-called AA (advection – aridity) model for calculating actual evapotranspiration proposed by Brutsaert and Stricker (1979), where evapotranspiration is calculated by collating information from energy balance and transfer of water vapour from Penman's equations (Xu and Chen 2005). After sorting, the actual evapotranspiration is expressed as:

$$ET_A^{AA}=(2\alpha-1)\frac{\Delta}{\Delta+\gamma}\frac{R_n}{\lambda}-\frac{\gamma}{\Delta+\lambda}f(U_2)(e_s-e_a) \tag{11}$$

where α is coefficient (α=1.26); Δ is gradient of saturated water vapour pressure (mbar/°C); γ *is* psychometric constant (mbar/°C); R_n is net radiation (cal/cm^{2}day); λ is latent heat of evaporation (cal/g); $f(U_2)$ is function of average wind speed at a height of 2 m above ground (m/s); e_s is air water vapour pressure (Pa); e_a is saturated water vapour pressure at air temperature (Pa).

Biftu and Xu cited the so-called GG model which was proposed by Granger and Gray (1989) which works by modifying Penman's equations for estimating actual evapotranspiration from different unsaturated ground covers (Biftu and Gan 2000; Xu and Singh 2005):

$$ET_A^{GG}=\frac{\Delta G}{\Delta G+\gamma}\frac{R_n}{\lambda}+\frac{\gamma \cdot G}{\Delta G+\gamma}E_a \tag{12}$$

where G is dimensionless relative evapotranspiration parameter; γ is psychometric constant (mbar/°C); R_n is net radiation in neighbouring area (cal/cm^{2}day); λ is latent heat (cal/g); E_a is drying power of air (mm/day).

Further, Xu cited the so-called CRAE model proposed by Morton (1978). He separated Penman's equations into two parts which he characterises as the energy balance and the water vapour transfer process (Xu and Chen 2005; Xu and Singh 2005). Refinement involves the introduction of 'balanced temperature':

$$ET_A^{CRAE} = 2\,ET_W^{CRAE} - ET_P^{CRAE} \tag{13}$$

$$ET_W^{CRAE} = b_1 + b_2 \frac{\Delta_p}{\Delta_p + \gamma} R_{Tp} \tag{14}$$

$$ET_P^{CRAE} = R_T - \left[\gamma f_T + 4\,\varepsilon\sigma\left(T_P + 273\right)^3\right]\left(T_P - T\right) \tag{15}$$

where ET_W^{CRAE} is wet environment evapotranspiration (W/mm^2); ET_P^{CRAE} is potential evapotranspiration (W/mm^2); b_1 represents the minimum of energy which serves for ET_W (b_1 = 14 W/m); b_2 replaces the Priestley-Taylor factor α (b_2 = 1.2); Δ_p is gradient of saturated water vapour pressure (mbar/°C); γ is psychometric constant (mbar/°C); R_{Tp} is net free energy (cal/cm^{2}day); R_T is net surface radiation at air temperature (cal/cm^{2}day); f_T is vapour transfer coefficient; ε is emissivity of area; σ is Stefan-Bolzmann constant; T_P is balanced temperature (°C); T is air temperature (°C).

Zhang worked out an algorithm which calculates monthly evapotranspiration on a basis of data collected by the help of satellites (Zhang et al., 2009). After classifying individual pixels by cover types, the calculation of evapotranspiration is carried out depending on ground cover type. Plant transpiration is calculated by the basic Penman-Monteith equation already described (4), while evaporation above water surfaces is calculated with the help of the Priestley-Taylor equation which states:

$$\lambda \cdot E_{WATER} = a \frac{\Delta \cdot A}{\Delta + \gamma} \tag{16}$$

where a is a constant (a=1.26); Δ is gradient of saturated water vapour pressure (Pa/K); A is available energy for evaporation (W/m^2); γ is psychometric constant (Pa/°C). Soil evaporation is calculated using the equation proposed by Mu (2007) as an additional development of the Penman-Monteith equation:

$$\lambda \cdot E_{SOIL} = RH^{(VPD/k)} \frac{\Delta A_{soil} + \rho \cdot C_p \cdot VPD / r_a}{\Delta + \gamma\, r_{totc} / r_a} \tag{17}$$

where RH is relative air humidity (values between 0 and 1); VPD is soil water vapour deficit (Pa); k is parameter (k=100 Pa); Δ is gradient of saturated water vapour pressure (Pa/K); A_{soil} is available energy for soil evaporation (W/m^2); ρ is air density (kg/m^3); C_p is specific air heat capacity (J/kgK); r_a is aerodynamic resistance (s/m); γ is psychometric constant (Pa/°C); r_{totc} is corrected value of term r_{tot}, which refers to total aerodynamic resistance of water vapour (s/m).

Xu compared seven models of evapotranspiration calculation, of which three were for actual evapotranspiration (AA model, GG model, CRAE model) and their performance in determining water balance (Xu and Chen, 2005). The results demonstrate that for the calculation of actual evapotranspiration, the GG model gave better results than the other models. For the calculation of groundwater recharge, the GG and AA models gave the best results. For the calculation of soil moisture, four of the seven models, (out of which only one is used for calculating actual evapotranspiration – the GG model), gave equally acceptable results. Xu concludes that water balance components (actual evapotranspiration, groundwater recharge and soil moisture) may be predicted with satisfactory accuracy by the use of the GG model (along with the Makkink model mentioned above).

In another paper, Xu compared the results of calculations of the three above-mentioned models of actual evapotranspiration (AA model, GG model, CRAE model) at three locations with very different geographical and climatological features (Sweden, eastern China and Cyprus). The comparison showed that for yearly time step all three models give satisfactory results, while for monthly time step in a dry climate (Cyprus) was somewhat worse. He states that the CRAE model showed slightly better accuracy than the other models (Xu and Singh, 2005).

Surface runoff

Among equations for water balance it is worth mentioning the frequently-used Turc equation (1954). Turc calculated runoff deficit as a function of rainfall and temperature (Horvat and Rubinic 2006):

$$D=\frac{P}{\sqrt{0.9+\frac{P^2}{L^2}}} \tag{18}$$

$$L=300+25T+0.05T^3 \tag{19}$$

where P is precipitation (mm); L is temperature factor; T *is* air temperature (°C).

For modelling water balance, Shen divides surface runoff into surface flow and streamflow. Surface flow is modelled using two-dimensional wave equations (Shen and Phanikumar, 2010):

$$\frac{\partial h}{\partial t}+\frac{\partial(h\cdot v)}{\partial x}+\frac{\partial(h\cdot u)}{\partial y}=s \tag{20}$$

$$0=-g\frac{\partial h}{\partial x}+g(S_{0x}-S_f) \tag{21}$$

$$0=-g\frac{\partial h}{\partial y}+g(S_{0y}-S_f) \tag{22}$$

where h is depth of surface water flow (m); u,v are flow velocity in x,y directions (m/s); s is the source runoff (m/s); S_0 is slope(-); S_f is friction slope (-). The modelling of streamflow is based on a one-dimensional wave equation:

$$\frac{\partial A_c}{\partial t}+\frac{\partial(uA_c)}{\partial x}=\frac{P}{86400}w+q_{oc}+q_{gc}+q_t \tag{23}$$

$$0=-g\frac{\partial h_r}{\partial x}(S_{0x}-S_f) \tag{24}$$

where A_c is surface cross-section (m^2); u is flow velocity (m/s); q_{oc} is lateral flow from surface flow (m^3/m/s); q_{gc} is contribution of subsurface water (m^3/m/s); q_t is tributary contribution (m^3/m/s); w is width of watercourses (m).

In the MIKE SHE software, the simplified equation for surface flow which is based on the continuity equation and Manning equation (DHI, 2007). The continuity equation states:

$$\frac{\partial q}{\partial x}=R-\frac{\partial y}{\partial t} \tag{25}$$

where q is specific flow (m^2/s); R is precipitation (mm); x is positive flow direction; y is local surface water depth (m). Manning's equation for turbulent flow may be described as:

$$q=M\cdot y^{5/3}\sqrt{\alpha} \tag{26}$$

where M is Manning's coefficient; α is the slope of surface terrain (-). After certain hypothesis and calculations of above equations and substitution in a continuity equation, which show that total volume of outflow is equal to total volume of runoff minus the change in volume at a soil surface, an equation of specific flow is obtained:

$$q=M\sqrt{y}\left[\frac{D}{L}\left(1+\frac{3}{5}\left(\frac{D}{De}\right)^3\right)\right]^{5/3} \tag{27}$$

where L is length of sloping part of watershed area (m); D is water retained on surface before balancing (m^3/m); De is water retained on surface (m^3/m).

Infiltration

Infiltration through soil surface is the link between surface and subsurface flow. From the many large source methods used to assess infiltration, three groups of methods may be singled out which are easy enough to use but provide estimates which have a scientific basis, experiential models, Green-Ampt models and models based on Richards' equation (Ravi and Williams, 1998; Williams et al., 1998). Among the experiential equations there are enumerated equations developed by Kostiankov (1932), Horton (1940), Mezencev (1957), Holtan (1961) and Broughton (1966). The empirical term proposed by the USDA Soil Conservation Service (1957) is often used:

$$R=\frac{(P-Ia)^2}{(P-Ia)+S} \tag{28}$$

where R is runoff (m^3/s); P is precipitation (mm); S is maximum retention capacity (mm); Ia is initial abstraction (mm). Infiltration is calculated as the difference between precipitation and runoff:

$$I=P-R \tag{29}$$

Green and Ampt (1911) developed the first equation based on process physics which described water infiltration in the ground. Their model experienced many improvements over time (e.g. Bouwen (1969), Childs and Bybordi (1969), Swartzendruber (1974), Chu (1978), Philip (1992) etc.). Due to the simplicity and satisfactory properties of this model it is often used, especially where the use of complex approaches (such as Richards' equations) is impractical, because of the need for a series of ground hydraulic parameters. Infiltration is defined as (Ravi and Williams, 1998; Williams et al., 1998; Ma et al., 2010):

$$i = K_s \frac{Z_f + H_0 + S_f}{Z_f} \tag{30}$$

where K_s is hydraulic conductivity (cm/min); Z_f is the wetting front depth (cm); S_f is the wetting front suction head (cm); H_0 is the depth of ponding water (m).

Water transfer between watercourses and surrounding saturated ground (aquifer) may be significant in permeable soils. Such sediment is defined in the MIKE SHE software package as the multiplication of permeability and the difference in potential between watercourse and ground (DHI, 2007):

$$Q = C \cdot \Delta h \tag{31}$$

where C is conductivity (m/s); Δh is potential difference (m). There are three different layers of transfer between watercourses and saturated ground: transfer only through the material of the aquifer, transfer only through riverbed material and transfer through both materials. In the case of transfer only through the material of aquifers, permeability is defined as:

$$C = \frac{K \cdot da \cdot dx}{ds} \tag{32}$$

where K is horizontal permeability (m/s); da is vertical area available for flow exchange (m^2); dx is cell size of saturated ground component (m^2); ds is average flow length (m). In the case of transfer only through riverbed material, permeability is defined as:

$$C = L_C \cdot w \cdot dx \tag{33}$$

where L_C is transfer coefficient for riverbed material (1/T); w is wetted perimeter of cross-section (m). In the case of transfer through the material of the aquifer and riverbed, permeability is:

$$C = \frac{1}{\dfrac{ds}{K \cdot da \cdot dx} + \dfrac{1}{L_C \cdot w \cdot dx}} \tag{34}$$

Subsurface flow

The calculation of subsurface flow is divided into the calculation of the unsaturated zone (above the underground water level) and the saturated zone (below the underground water level). With flow in the unsaturated zone the force which impels the water is the hydraulic pressure gradient (DHI, 2007):

$$h = z + \psi \tag{35}$$

where z is gravitational component (height); ψ is pressure component. Vertical flow is driven due to the vertical gradient of hydraulic pressure. Volumetric transfer is calculated by Darcy's law:

$$q = -K(\theta) \frac{\partial h}{\partial z} \tag{36}$$

where $K(\theta)$ is unsaturated hydraulic conductivity and $\partial h/\partial z$ is vertical gradient of hydraulic pressure. Ground water flow in the unsaturated zone is described by Richards' equation (Shen and Phanikumar, 2010; DHI, 2007; Williams et al., 1998):

$$C(h)\frac{\partial h}{\partial t}=\frac{\partial}{\partial z}\left[K(h)\left(\frac{\partial h}{\partial z}+1\right)\right]+W(h) \tag{37}$$

where $C(h)$ is differential water capacity; $K(h)$ is unsaturated hydraulic conductivity; $W(h)$ is sink volume (including contribution of evaporation and plant root extraction).

Flow in the saturated zone is described by a finite differential method with the ground divided into a series of layers in which flow is described in two-dimensional or three-dimensional flow equations (DHI, 2007).

WATER BALANCE NUMERICAL MODELS

Many researchers have developed numerical models for calculating water balance which have different levels of complexity. This is primarily related to differences in the quantity of required input data, the range of calculated output data and their applicability to areas of different sizes. It is required to balance the needs and expectations of end users with the possibilities of single models and with frequent restrictions in availability and reliability of geophysical parameter measurements (hydrological, hydraulic, geodesic, hydrometeorological, hydrogeological, etc.).

Alemaw developed the DGHM model (Distributed GIS-based Hydrological Model) as a mean of documenting seasonal regimes of ground moisture, actual evapotranspiration and runoff and to present them in geographically referenced patterns on a continental scale (Alemaw and Chaoka, 2003). Evapotranspiration was estimated on the basis of modified Thornthwaite equation (2), surface runoff by the SCS method (28) and soil moisture on the basis of relation between total precipitation and potential evapotranspiration.

Sanchez developed the spatially distributed model HIDROMORE which uses a calculation approach to water balance which utilizes the FAO-56 method and provides daily values of hydrological parameters (infiltration, storage of water and evapotranspiration) for the entire watershed area (Sanchez et al., 2009). For estimating evapotranspiration they used the Penman-Monteith equation and some of required variables were estimated by remote sensing.

For estimating the hydrological water balance and water requirement for irrigation, Portughese developed a model based on GIS which comprises an analytical model for estimating soil moisture and a unified model for estimating the level of underground water (Portughese et al., 2005). Separation of total monthly precipitation into net infiltration and surface runoff was calculated on the basis of the SCS method (28) while evapotranspiration was estimated using the Penman-Monteith equation (5). The model was developed for the needs of an area covering the whole region comprising several watersheds.

For long-term simulations on watersheds of average (∼1000 km^2) and large (>5000 km^2) surfaces, Shen developed the distributed hydrological model PAWS (Process-based Adaptive Watershed Simulator). For estimating evapotranspiration it uses the Penman-Monteith equation (5). Surface runoff is calculated through two-dimensional (20) (21) (22) and one-dimensional wave equations (23) (24). Infiltration is estimated using Richards' equation (37) while for heavy precipitation the Green-Ampt model (30) is used. Underground flow in the unsaturated zone is calculated using Richards' equation (37) and flow in the saturated zone is calculated using two-dimensional flow equations. Special care is devoted to the interaction between the described components (Shen and Phanikumar, 2010).

Previously demonstrated models (Alemaw and Chaoka, 2003; Sanchez et al., 2009; Portughese et al., 2005; Shen and Phanikumar, 2010) use only one approach for estimating water balance components depending on the availability of input data. There are software packages which allow users a choice of approach for estimating water balance components and are adjusted with different kinds of input data and also to domains of different sizes. An example of such a model is the conceptual watershed model GEOTRANSF (AquaTerra Project, 2005). Its structure and the methods used for calculating water balance components are shown graphically (Figure 1).

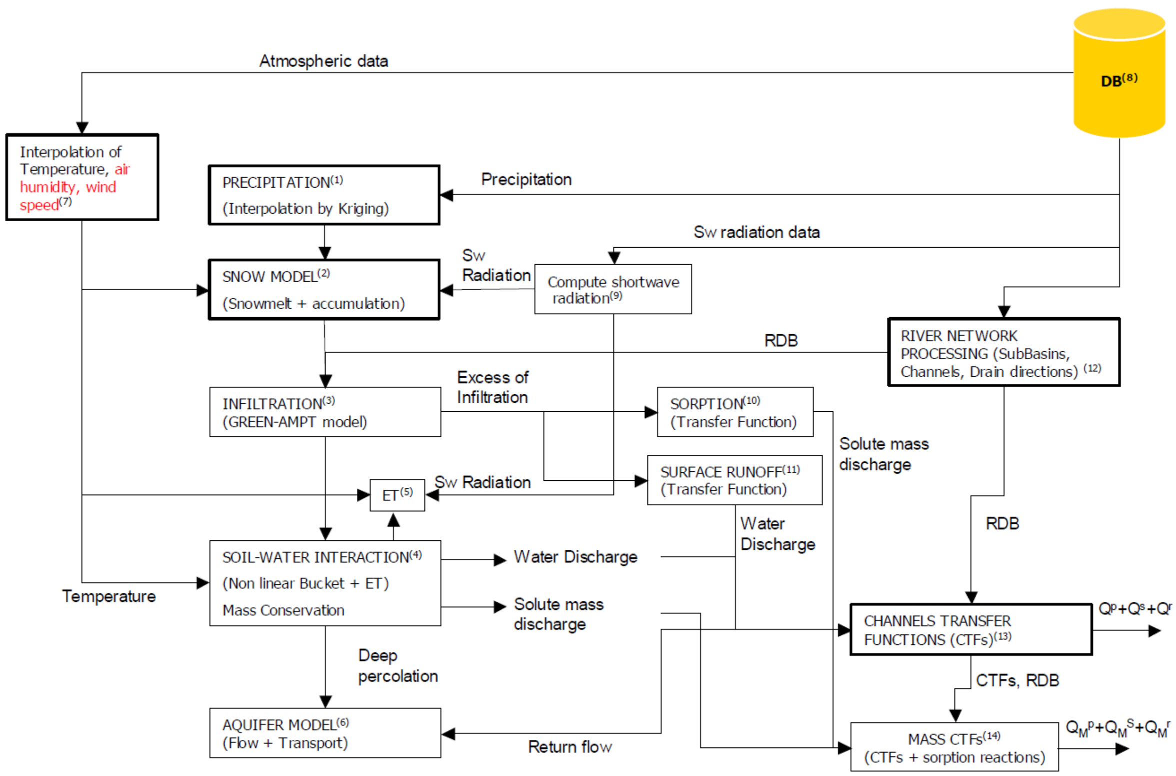

Figure 1. GEOTRANSF model structure (AquaTerra Project, 2005)

The completely integrated distributed model MIKE SHE is another example of such a model (DHI, 2007). This software package enables simulations of a large number of hydrological and hydraulic water balance components. It is applied to watersheds of different sizes in moist and dry climates and also uses spatially distributed continual climate data. Figure 2 shows the computational structure of the MIKE SHE model and available procedures for calculating individual water balance components.

EXAMPLES OF APPLICATIONS OF WATER BALANCE MODELS

Along with a review of different model structures and approaches to the calculation of balance components, additionally some application examples of models for different spatial and time domains are shown. The DGHM water balance model (Alemaw and Chaoka, 2003) is used in areas of Southern Africa particularly in SADC region (the region between 0°-35° S and 5°-55° E) within the period 1961 to 1990 with the help of GIS variations in soil moisture, actual evapotranspiration and runoff on a 0.5° x 0.5° grid.

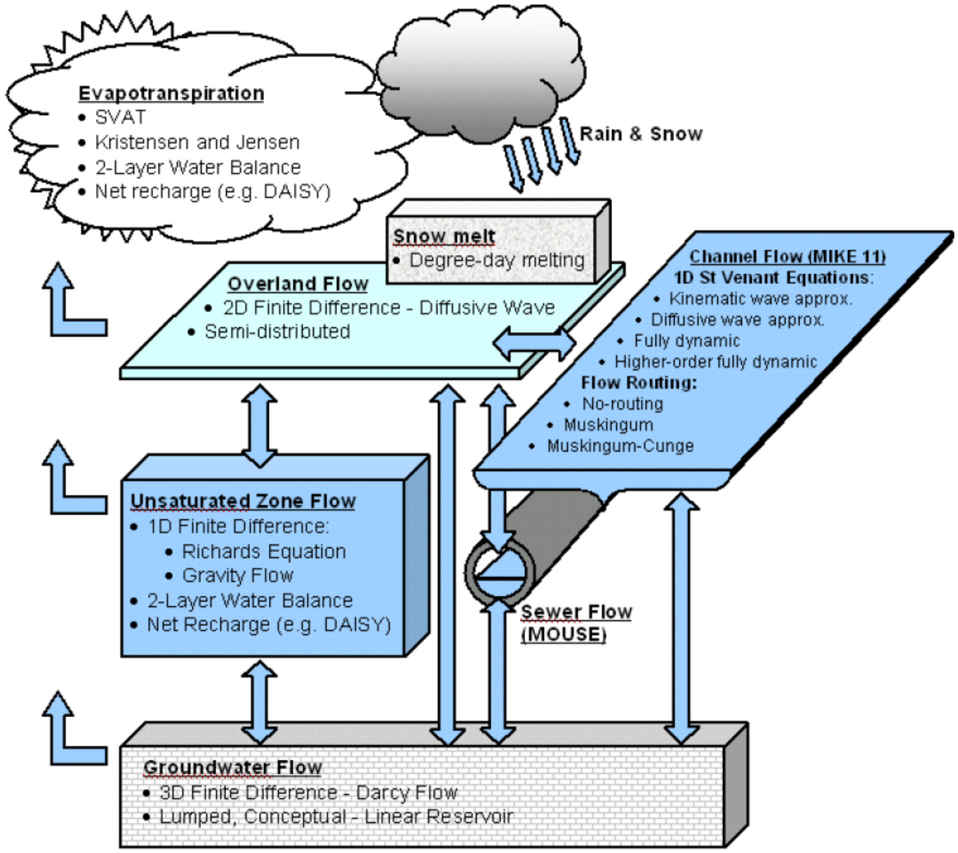

Figure 2. Schematic display of MIKE SHE model structure (DHI, 2007)

With the use of the HIDROMORE model (Sanchez et al., 2009) using a computational network resolution from 3kmx3km (total 146 cells) the water balance was calculated in the 1300km² watershed of the river Duero in Spain. For estimating evapotranspiration, Landsat satellite recordings were used. Model calibration and validation were conducted for the period of 1 year by comparing calculated and measured moisture values in 23 measuring stations with continual measurements of soil moisture. Although the model, which focussed primarily on the calculation of soil moisture, yielded soil moisture values somewhat less than measured, it demonstrated effectiveness in the studied watershed.

For calculating the water balance in the 13000 km² Italian Puglia region the Portughese model was used with a spatial resolution of 1km x 1km (Portughese et al., 2005). From available climate data of a 40-year period, the water balance was calculated with a time step of 1 month and the calculated values of underground water levels were equal to measured levels in wells. The model proved very adaptable and gave satisfactory results for underground water recharge, needs for irrigation water and soil moisture. Furthermore, the model offers estimates of crop water requirements with regard to different climatic and control scenarios.

The GEOTRANSF model (AquaTerra Project, 2005) has been thoroughly tested in multiple watersheds, one example of which is the calculation of the water balance in the Ebro river basin in Spain. For the calculations, measured data of precipitation, air temperature, potential evapotranspiration and flow measurement in a series of stations were available as well as a digital ground model of 90m x 90m spatial resolution. Model calibration and validation were executed for the control period 1961 to 1990 in 4 sub-catchments (marked in yellow in Figure 3).

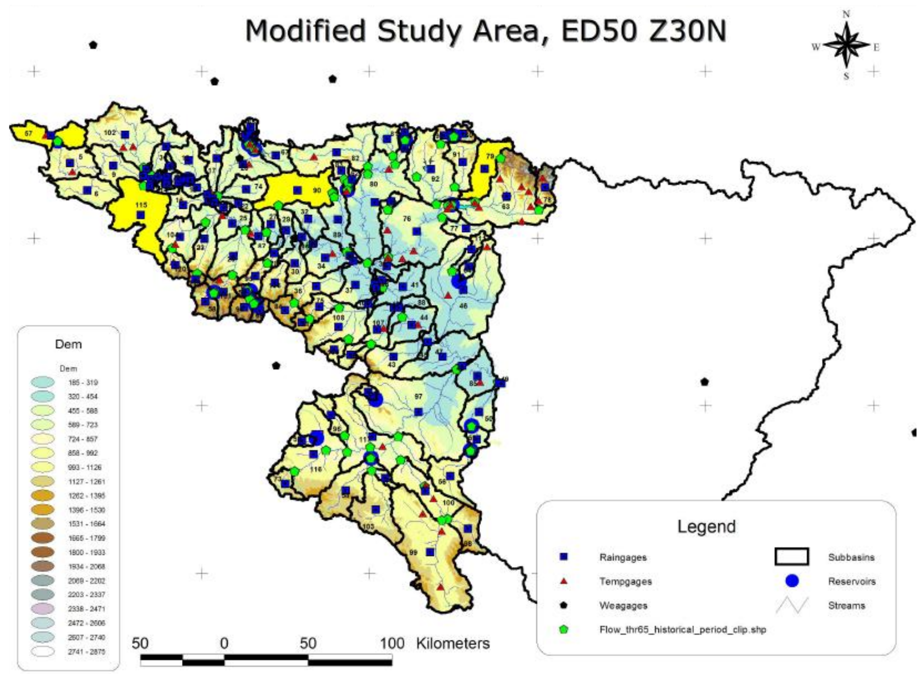

Figure 3. The River Ebro basin in which GEOTRANSF was tested (AquaTerra Project, 2005)

Figure 4 shows a comparison of measured and calculated monthly flow during the model calibration and validation. In the model estimates of runoff in the sub-catchments for the period 2071 to 2100 are also given, with regard to expected variations in precipitation and air temperature.

Widen-Nilsson developed the global water balance model WASMOD-M (Water and Snow balance MODelling system – Macro-scale) which offers general estimates of runoff (Widen-Nilsson et al., 2007). With the resolution of 59132 computational cells of 0.5°x0.5° they simulated the worldwide water balance for the period 1915-2000 by using a time step of 1 month within 14-year warm-up period (1901-1915). It demonstrates estimates of runoff within ±20% from measured values for 455 out of 663 measuring stations and within ±1% for 276 measuring stations.

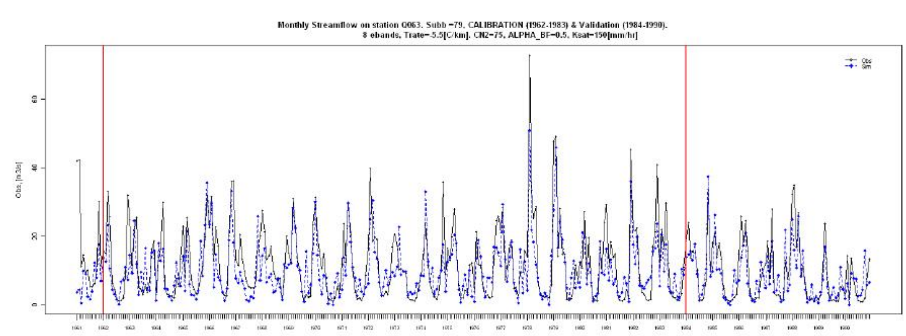

Figure 4. Comparison of measured and calculated averaged monthly discharges for calibration (1962-1983) and validation (1984-1990) (AquaTerra Project, 2005)

Calibration and validation of the integrated hydrological model MIKE SHE is shown for the island of Sjælland in Denmark whose area of 7330 km² was described by 1km x 1km computational cells (Henriksen et al., 2003). The model was produced with the use of national data for geology, pedology, topography, watercourse networks, climate and hydrology. Calibration (1988-1990) and validation (1991-1996) of the model was conducted on the basis of measurements of the underground water level in 4439 wells and water levels in 28 stations (Figure 5). Figure 6 shows model validation example for one water measuring stations. The model produced reliable estimate of underground water levels and recharge of various geological layers with subsurface water considering different variations of input climate parameters possible.

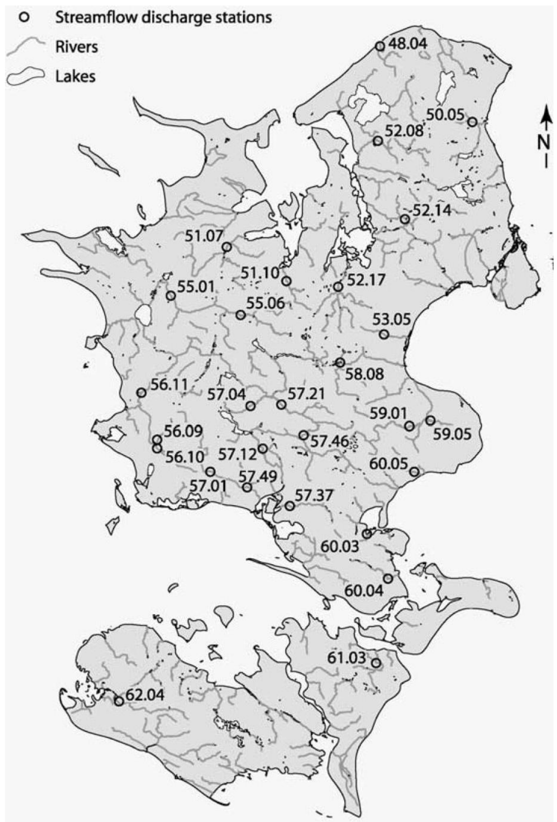

Figure 5. Sjælland Island and 28 water measuring stations for the calibration and validation of the MIKE SHE model (Henriksen et al., 2003)

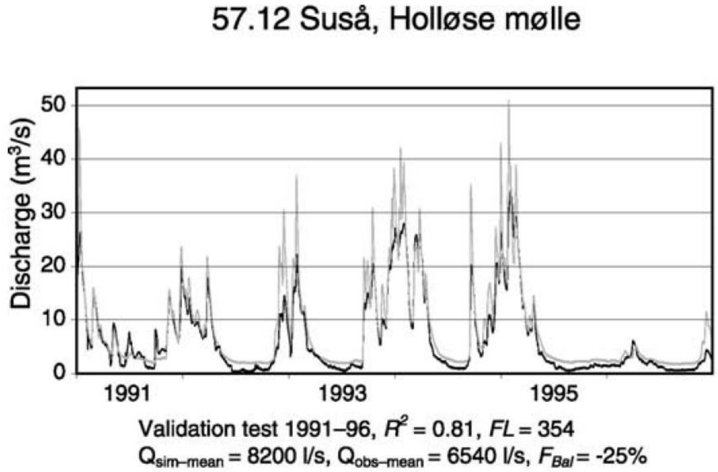

Figure 6. Comparison of calculated and measured flow at one water measuring stations (1991-1996) (Henriksen et al., 2003)

CONCLUSION

Climate changes and its negative consequences are observed on each continent. Current data for Europe shows that at least 11% of inhabitants and 17% of land area are threatened by lack of water, and that drought-related damage amounted to 100 billion Euros during the last 30 years (EEA, 2009). Estimates of climate changes show an increase in the frequency of extremely hot periods as well as heavy rainfall on the global level (IPCC, 2007). Although changes in precipitation and evaporation are predicted with relatively less reliability in climate models (Boughton, 2005), predictions for the European continent show precipitation increasing in the north and decreasing in the south, with consequent direct influences on regime changes of surface and subsurface water on a wider scale (Xu, 1999) and indirect influences on changes in water availability, on loss of biodiversity and on other outcomes. Negative consequences of climate change and the curtailment of its associated risks may be alleviated by implementing different measures as well as simply adjusting to change. In this context water balance calculations have a significant role.

Water balance is analysed on a wide spectrum of spatial domains from 'small' watersheds up to whole regions and in time periods ranging from 1 year to several decades. According to the case analysis discretisation of spatial domains is carried out with spatial increments from 1 metre to some tens of kilometres and discretisation of time domain with time steps from 1 day to 1 month. For a reliable description of the complex processes of runoff and nonlinear links between every component it is necessary to use complex mathematical models and to define detailed spatial and temporal discretisation of the problem. The requirement for additional parameterisation processes (Beven, 2006) and for a model of detailed spatial resolution and short-term time step necessitates exceptional computing capacity. For the reliability of water balance models, a continuously measured series of geophysical parameters over a wide area are equally important.

Although various empirical procedures and methods are currently available, the main deficiency in reliable water balance estimates lies in the absence of long-term continual hydrometeorological measurements and incomplete knowledge about features of watersheds (from geodesic, geological and pedological bases to data concerning vegetation, albedo (reflection coefficient), surface temperatures, ground moisture and etc.). Information about the variability of hydrological size in space and time may be acquired on the basis of in-point measurements along with subsequent spatial interpolations or extrapolations (Horvat, 2012). High spatial and temporal resolution satellite technology is becoming increasingly important source of data for observing hydrological parameters in larger areas. New satellite images (Eumetsat and Landsat satellites) provide visible spectrum data, water vapour spectrum data and infrared spectrum data with which data concerning albedo, vegetation index and surface temperatures are obtained and indirectly about net radiation, soil heat energy and energy transferred to the atmosphere, etc. Estimating evapotranspiration and soil moisture with remote sensing are also areas of rapid development (Immerzeel and Droogers, 2008).

For the purposes of estimating the water balance in developed countries, precipitation, air, ground and water temperature, air humidity, insolation and wind speed are measured at meteorological stations, and water levels and watercourse flows are systematically measured at hydrological stations. A national digital terrain model may be obtained from the State Geodetic Administration. Cover features are available via the Corine Land Cover Atlas. Further, Geological and Hydrologeological Maps are usually available. Consistent detailed research into geological structure and the tracking of underground water levels at a national level on the other hand are seldom available. Measurement work is often carried out only in specified time periods and in association with specific projects. From this overview of contemporary procedures for calculating water balance it may be concluded that real balance results may be obtained primarily only with the use of reliable input data. In other words, the reliability of results will primarily depend on the quality and reliability of observations on climatic and hydrological parameters and less on the selected calculation model. For reliable estimates and research of availability of water on the local and global level, the need for continual monitoring of hydrological parameters has already been highlighted (Geres, 2004).

In the selection of individual approaches and models, as well as assessing the technical advantages and limitations of individual programming packages, it is necessary to perceive the sustainability of the 'system'. Sustainability of individual packages is manifested in the possibility of implementing new methods and approaches to define individual hydrological processes but even more significantly in model validation through longer time periods and in the possibilities of 'system learning' (Sanchez et al., 2010). Furthermore, commercial programming and 'open-source' packages have been developed which are available to a wide spectrum of users. 'Open source' packages do not require initial investment but give greater risks with package updates which cannot be relied on in the future. Therefore it may be freely stated that the selection of individual approaches and models for specific area is an important determinant for the individual as much as for the organisation involved.

Given that water balance model results are used for different purposes and end-users, in this selection of modelling approaches for specific cases it is therefore necessary to take into consideration the features of software packages, the availability of input data, needs of end-users and sustainability of software packages. Climate variation may change water balance significantly in the form of both the increased peak flows in rivers (which will increase flood risk, scour risk at bridges, etc.) and the reduced low flows and increased drought periods (which will influence water supply, agricultural production, etc). It must be concluded that the selection of water balance models is also one of the strategic milestones which will enable scientists, various key parties in water management planning and water users to estimate effects of climate variation on water resources, to spot weaknesses in water availability and define adjustment measures according to climate change in their future planning.

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