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Large Scale Groundwater Modelling

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Abstract

The hydrological model LISFLOOD is a spatially-distributed, partially physically-based rainfall-runoff model that has been devised to simulate the hydrological behaviour in large European catchments, with emphasis on predicting floods and droughts. In the model the groundwater component is represented by two interconnected linear reservoirs, with the outflow in some unit of time being proportional to the volume of water stored in the reservoirs. Several studies have shown that this setup of the model is able to reproduce the observed hydrographs reasonably well, hence the model can be considered to be sufficiently detailed to predict high (floods) and low (droughts) flows. However, LISFLOOD is not able to simulate the spatial distribution of groundwater levels and a comparison with groundwater observations is not possible. This document provides an overview on large-scale groundwater modelling techniques, and proposes possible ways to adapt LISFLOOD such that it can provide an estimate of the groundwater elevation in space and time.

Content

1. Introduction

The goal of this report is to evaluate large scale groundwater modelling techniques to be used at the European scale for flood and drought forecasting and hydrological impact assessment studies. In the current version of the LISFLOOD model, the groundwater component is represented by two interconnected linear reservoirs, with the outflow in some unit of time being proportional to the volume of water stored in the reservoirs. Simulations with this setup have shown to yield acceptable reproductions of the observed hydrograph, hence the model can be considered to be sufficiently detailed to predict high (floods) and low (droughts) flows. However, a good reproduction of river discharges does not imply that other hydrological variables (soil moisture, groundwater levels) or processes (percolation, plant water uptake, and retention processes) are well reproduced.

LISFLOOD is not able to simulate the spatial distribution of groundwater levels, hence a comparison with groundwater observations is not possible. This document proposes possible ways to adapt LISFLOOD such that it can provide an estimate of the groundwater elevation in space and time. This would also allow distributed groundwater measurements to be used in the calibration and validation of the model, and to gain insight in the local hydrological processes and their representation in the model.

The document is organised as follows. Section 2 presents some general information and principles about groundwater. A general description of groundwater models is provided in Section 3. In Section 4 an overview of the representation of groundwater in regional scale catchment models is presented. Details about the current setup of the groundwater component in LISFOOD are given in Section 5. Two possible ways to adapt the groundwater module of LISFLOOD are presented in Section 6, and conclusions and suggestions for the way forward can be found in Section 7.

2. General information on groundwater

2.1. Groundwater and the hydrological cycle

The continuous circulation of water as it moves in its various phases through the atmosphere to the earth, over and through the land, to the ocean, and back to the atmosphere, is known as the hydrological cycle. In this work we exclusively consider the land phase of the hydrologic cycle. The different parts of the land-based portion of the hydrological cycle that take place at the individual watershed or catchment scale are represented schematically in Figure 2.1. Water enters the hydrological system as precipitation, in the form of rainfall or snowmelt. Precipitation is delivered to streams on the land surface as overland flow to tributary channels and in the subsurface as interflow or lateral subsurface flow and base flow following infiltration into the soil. Water leaves the system as stream flow or runoff, as deep groundwater flow crossing the catchment boundary, and as evapotranspiration, a combination of evaporation from open bodies of water, evaporation from soil surfaces, and transpiration from the soil by plants.

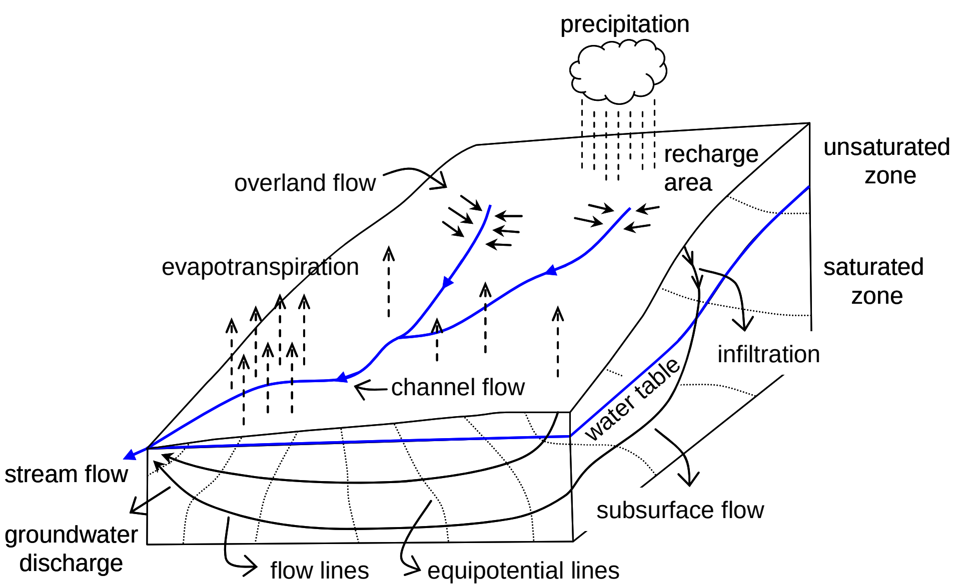

Figure 2.1. Schematic representation of the land phase of the hydrological cycle (adapted from Freeze, 1974).

The portion of infiltrated water that percolates through the vadose or unsaturated zone and enters the saturated zone or aquifer system is called groundwater. Recharge by seepage from surface water bodies can also replenish groundwater resources. Water leaves groundwater storage primarily by discharge into rivers or lakes, but it is also possible that water moves upward from the water table into the capillary fringes.

2.2. Principal controls on groundwater flow

Groundwater possesses energy mainly by virtue of its elevation (elevation or gravitational head) and of its pressure (pressure head). Kinetic energy is negligible because of the low velocities of groundwater when moving through the subsurface. The hydraulic head, or the height that water stands above a reference datum, is a measure of the energy of groundwater. The change in hydraulic head over a certain distance, termed hydraulic gradient, constitutes the main driving force for groundwater movement. It is predominantly determined by gravity forces due to elevation differences. Hubbart [1940] showed that, given an areally uniform precipitation and infiltration rate over an undulating surface, a groundwater flow system will develop driven by a water table surface that is a subdued replica of the land surface. Therefore, groundwater moves from interstream (higher) areas toward streams, lakes or the coast (lower areas), and the depth to the water table is generally greater along the divide between streams than it is beneath floodplains.

However, a groundwater flow pattern is not solely controlled by the configuration of the water table. It is also influenced by the distribution of the hydraulic conductivity in the rocks. Regions of high conductivity are made up of coarse-grained material with a large percentage of macropores that constitute areas that easily transmit water. The regions of low conductivity comprise fine-grained particles with a large percentage of meso- and micropores that restrict the rate of water movement. In addition to topographic and geological effects, groundwater flow is also affected by climate, precipitation being the source of recharge.

2.3. Groundwater flow systems

Groundwater moves along flow paths from areas of recharge to areas of discharge. Flow paths vary considerably in length, depth and travel time (ranging from days to millennia), and are organized in space to form a flow system. Subsurface flow domains can contain multiple flow systems of different orders of magnitude and relative, nested hierarchical order [Sophocleous, 2002]. Based on their relative position in space, Tóth

[1963] recognizes three distinct types of flow systems – local, intermediate and regional – which could be superimposed on one another within the groundwater basin. The different flow systems are presented in Figure 2.2. Water in a local flow system flows to nearby discharge areas, such as ponds or streams. Water in a regional flow system travels a greater distance and often discharges to major rivers, large lakes, or to oceans. An intermediate flow system is characterized by one or more topographic highs and lows located between its recharge and discharge areas, but, unlike the regional system does not occupy both the major topographic high and the bottom of the basin. Regional flow system are at the top of the hierarchical organization, with all other flow systems nested within them. Areas of pronounced topographic relief tend to have dominant local flow systems, whereas areas of nearly flat relief tend to be dominated by intermediate and regional flow systems [Sophocleous, 2004].

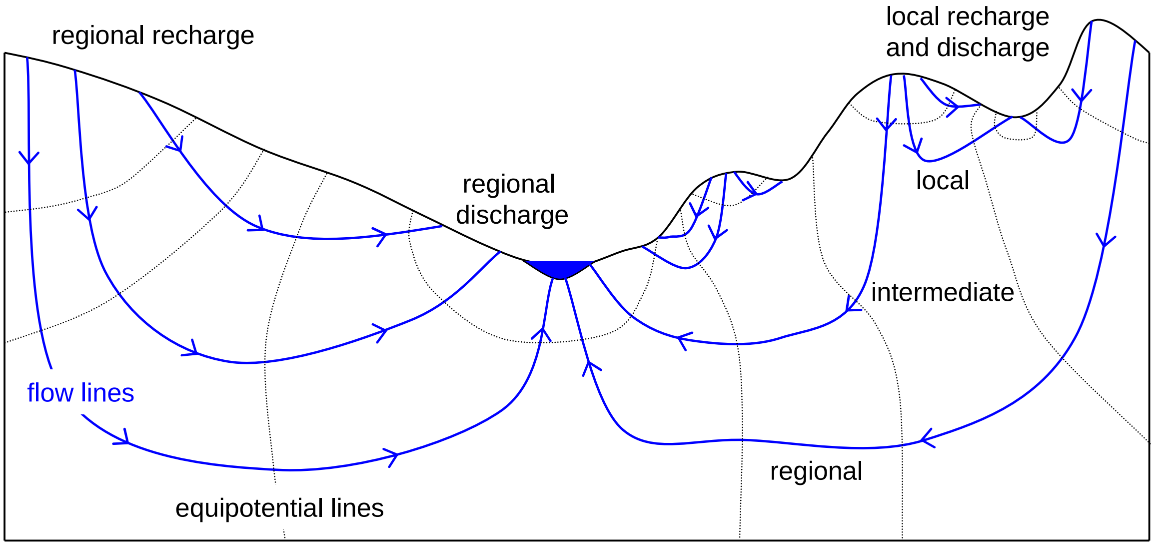

Figure 2.2. Groundwater flow systems in a regionally unconfined aquifer system (adapted from Tóth, 1999).

2.4. Interaction of groundwater and surface water

Groundwater interacts with surface water in streams, open water bodies and wetlands in a variety of ways. Because our interest is the forecasting of floods, or (extreme) levels of stream discharge, the discussion in this section will be limited to groundwater-stream interaction. Streams interact with groundwater in three basic ways: streams gain water from inflow of groundwater through the streambed (gaining stream, Figure 2.3), they lose water to groundwater by outflow through the streambed (losing stream, Figure 2.4), or they do both, gaining in some reaches and losing in other reaches. For groundwater to discharge into a stream channel, the water table elevation in the vicinity of the stream must be higher than the level of the stream water surface. Conversely, for surface water to seep to groundwater, the water table elevation in the vicinity of the stream must be lower than the stream water stage. The flow direction can vary a great deal along a stream. In some reaches the stream may receive groundwater, and in other reaches lose water to groundwater. Furthermore, the direction of flow can change within very short timeframes as a result of individual storms causing focused recharge near the stream bank, temporary flood peaks moving down the channel, or transpiration of groundwater by streamside vegetation [Winter et al., 1998].

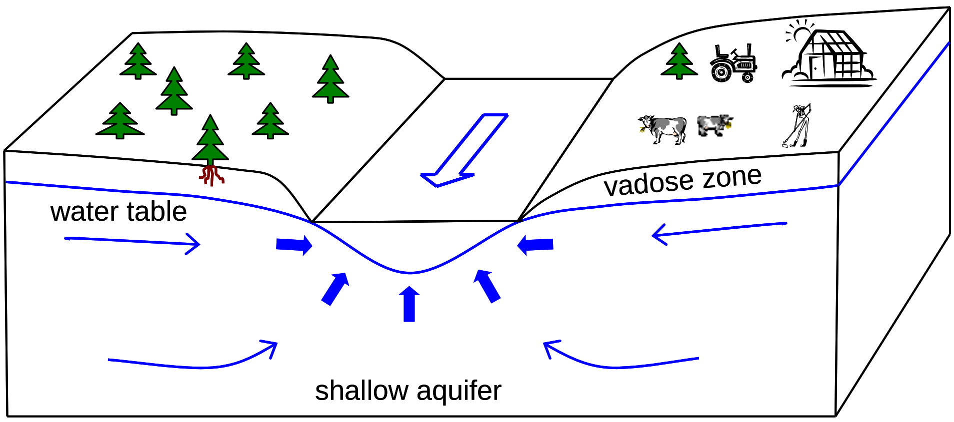

Figure 2.3. Interaction of groundwater and streams: a gaining stream (adapted from Winter et al., 1998).

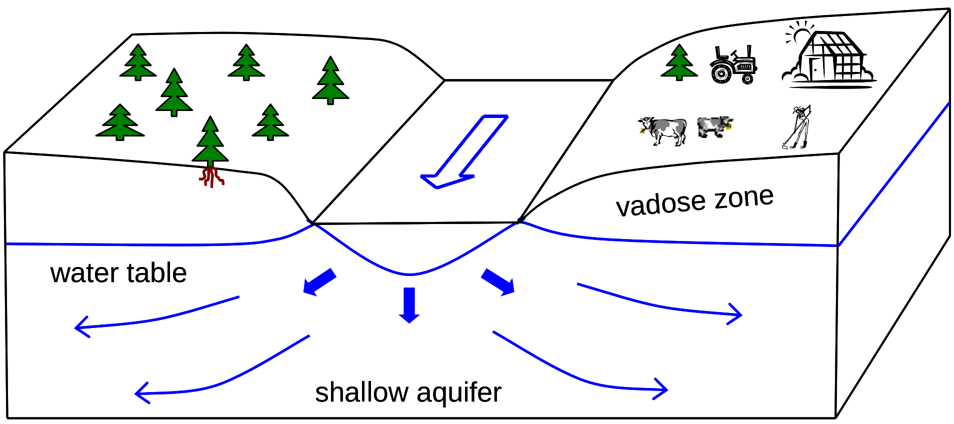

Figure 2.4. Interaction of groundwater and streams: a losing stream (adapted from Winter et al., 1998).

Losing streams can be connected to the groundwater system by a continuous saturated zone (Figure 2.4) or can be disconnected from the groundwater system by an unsaturated zone (Figure 2.5).

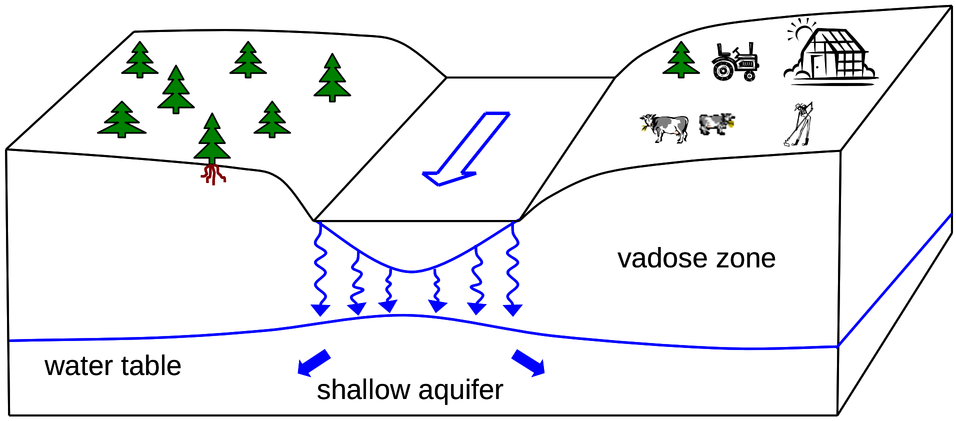

Figure 2.5. A disconnected stream separated from the groundwater system by an unsaturated zone (adapted from Winter et al., 1998).

Rapid rises in stream stage can cause water to move from the stream into the stream banks. This process, termed bank storage (Figure 2.6), is usually caused by storm precipitation, rapid snowmelt, or the release of water from a reservoir upstream. As long as the rise in stage does not overtop the stream banks, most of the volume of stream water that enters the stream banks returns to the stream within a few days or weeks. The loss of stream water to bank storage and return of this water to the stream in a period of days or weeks tends to reduce flood peaks and later supplement stream flows. If the rise in stream stage is sufficient to overtop the banks and flood large areas of the land surface, widespread recharge to the water table can take place throughout the flooded area.

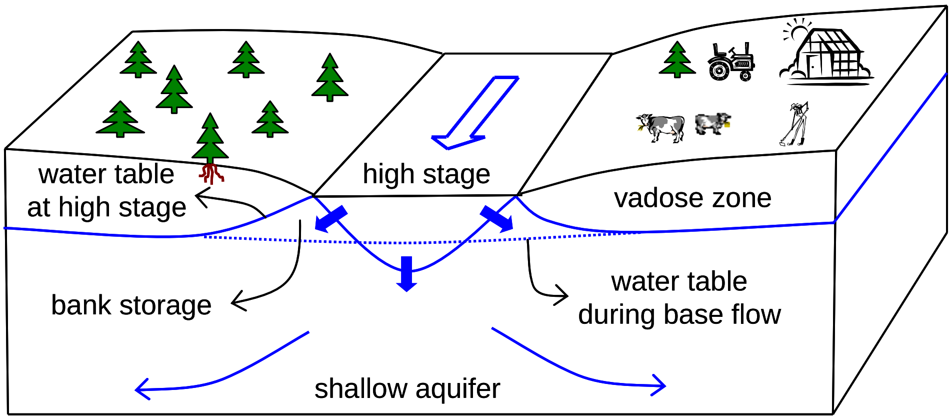

Figure 2.6. If stream levels rise higher than adjacent groundwater levels, stream water moves into the stream banks as bank storage (adapted from Winter et al., 1998).

2.5. Interaction of groundwater and surface water in different landscapes

The interaction of groundwater with surface water depends on the geomorphologic, geologic and climatic settings the of the landscape. For example, a stream in a wet climate might receive groundwater inflow, but a stream in an identical geomorphological and geological setting in an arid climate might lose water to groundwater. Some common features of the interaction for various parts of the conceptual landscape are described below. The five general types of terrain discussed are mountainous, riverine, coastal, glacial and dune, and karst.

2.5.1. Mountainous terrain

In mountainous terrain precipitation is highly variable and water moves over and through steep land slopes. On mountain slopes, macropores created by burrowing organisms and by decay of plant roots have the capacity to transmit subsurface flow downslope quickly. In addition, some rock types underlying soils may be highly weathered or fractured and may transmit significant additional amounts of flow through the subsurface. In some settings this rapid flow of water results in hillside springs.

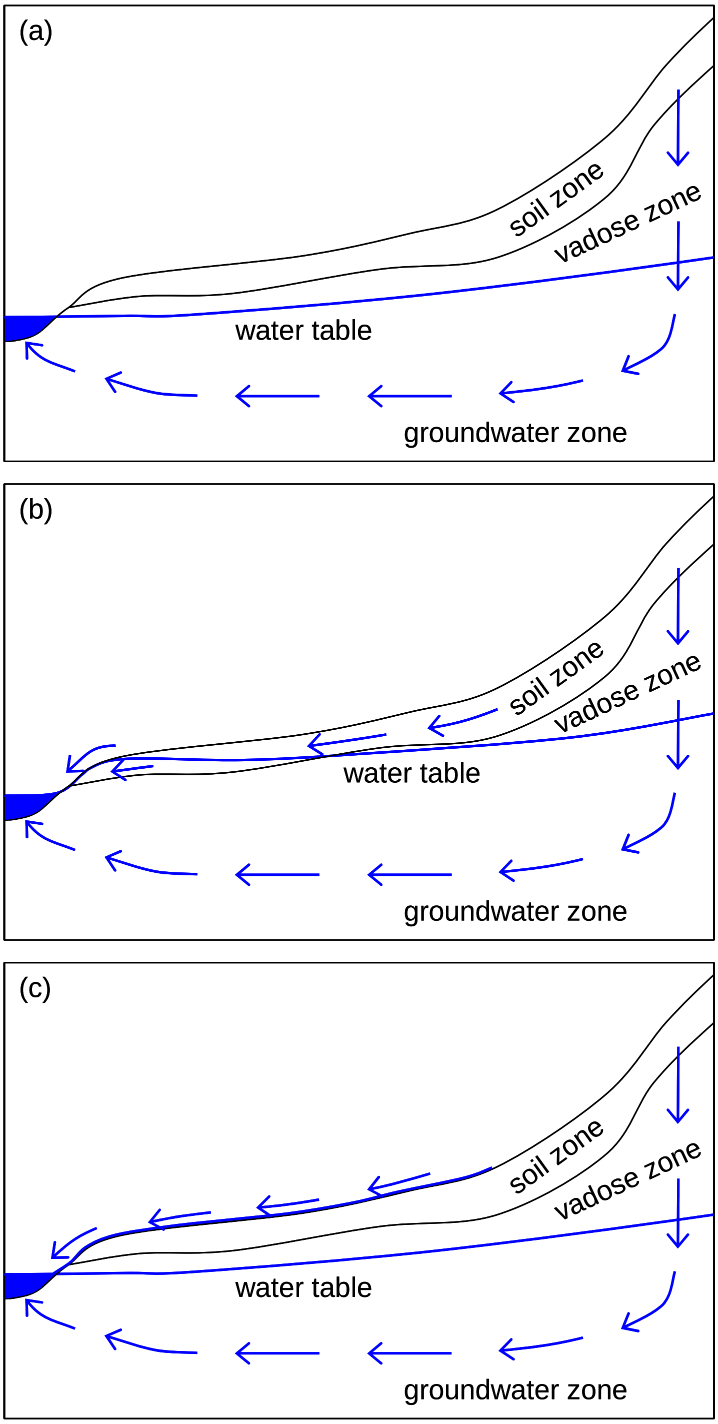

Figure 2.7. Flow of water in mountainous terrain (adapted from Winter et al., 1998).

A general concept of water flow in mountainous terrain includes several pathways by which precipitation moves through the hillside to a stream (see Figure 2.7). Between storm and snowmelt periods, stream flow is sustained by discharge from the groundwater system (Figure 2.7a). During intense storms, most water reaches streams very rapidly by partially saturating and flowing through the highly conductive soils. On the lower parts of hillslopes, the water table sometimes rises to the land surface during storms, resulting in overland flow (Figure 2.7b). When this occurs, precipitation on the saturated area adds to the quantity of overland flow. When storms or snowmelt persist in mountainous areas, near-stream saturated areas can expand outward from streams to include areas higher on the hillslope. In some settings, especially in arid regions, overland flow can be generated when the rate of rainfall exceeds the infiltration capacity of the soil (Figure 2.7c).

2.5.2. Riverine terrain

In some landscapes, stream valleys are small and they commonly do not have well-developed floodplains. However, major rivers have valleys that usually become increasingly wider downstream. Terraces, natural levees, and abandoned river meanders are common landscape features in major river valleys, and wetlands and lakes commonly are associated with these features.

The interaction of groundwater and surface water in river valleys is affected by the interchange of local and regional groundwater flow systems with the rivers and by flooding and evapotranspiration. Small streams receive groundwater inflow primarily from local flow systems, which usually have limited extent and are highly variable seasonally. Therefore, it is not unusual for small streams to have gaining or losing reaches that change seasonally.

For larger rivers that flow in alluvial valleys, the interaction of groundwater and surface water usually is more spatially diverse than it is for smaller streams. Groundwater from regional flow systems discharges to the river as well as at various places across the flood plain (Figure 2.8). If terraces are present in the alluvial valley, local groundwater flow systems may be associated with each terrace, and lakes and wetlands may be formed because of this source of groundwater. At some locations, such as at the valley wall and at the river, local and regional groundwater flow systems may discharge in close proximity.

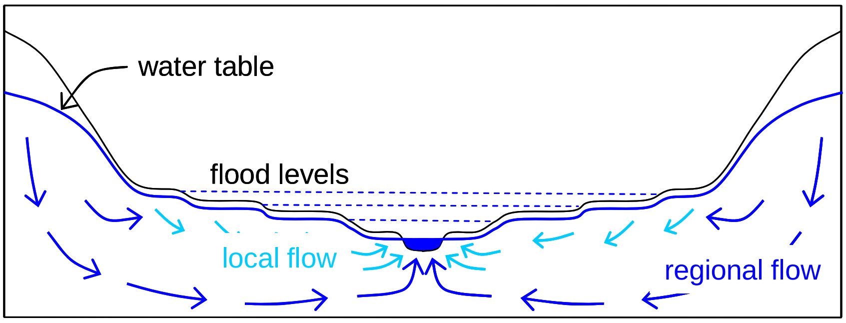

Figure 2.8. In broad river valleys, small local groundwater flow systems associated with terraces overlie more regional groundwater flow systems. Recharge from flood waters superimposed on these groundwater flow systems further complicates the hydrology of river valleys (adapted from Winter et al., 1998).

Added to this distribution of groundwater discharge from different flow systems to different parts of the valley is the effect of flooding. At times of high river flows, water moves into the groundwater system as bank storage (Figure 2.6). The flow paths can be as lateral flow through the riverbank or, during flooding, as vertical seepage over the floodplain. As flood waters rise, they cause bank storage to move into higher and higher terraces.

The water table generally is not far below the land surface in alluvial valleys. Therefore, vegetation on floodplains, as well as at the base of some terraces, commonly has root systems deep enough so that the plants can transpire water directly from groundwater. Because of the relatively stable source of groundwater, particularly in areas of groundwater discharge, the vegetation can transpire water near the maximum potential transpiration rate, resulting in the same effect as if the water were being pumped by a well. This large loss of water can result in drawdown of the water table such that the plants intercept some of the water that would otherwise flow to the river, wetland, or lake. Furthermore, in some settings it is not uncommon during the growing season for the pumping effect of transpiration to be significant enough that surface water moves into the subsurface to replenish the transpired groundwater.

2.5.3. Coastal terrain

Coastal terrain extends from inland scarps and terraces to the ocean. This terrain is characterized by (1) low scarps and terraces that were formed when the ocean was higher than at present; (2) streams, estuaries, and lagoons that are affected by tides; (3) ponds that are commonly associated with coastal sand dunes; and (4) barrier islands. Wetlands cover extensive areas in some coastal terrains.

The interaction of groundwater and surface water in coastal terrain is affected by discharge of groundwater from regional flow systems and from local flow systems associated with scarps and terraces (Figure 2.9), evapotranspiration, and tidal flooding. The local flow systems associated with scarps and terraces are caused by the configuration of the water table near these features. Where the water table has a downward break in slope near the top of scarps and terraces, downward components of groundwater flow are present; where the water table has an upward break in slope near the base of these features, upward components of groundwater flow are present. Evapotranspiration directly from groundwater is widespread in coastal terrain. The land surface is flat and the water table generally is close to land surface. Therefore, many plants have root systems deep enough to transpire groundwater at nearly the maximum potential rate. The result is that evapotranspiration causes a significant water loss, which affects the configuration of groundwater flow systems as well as how groundwater interacts with surface water.

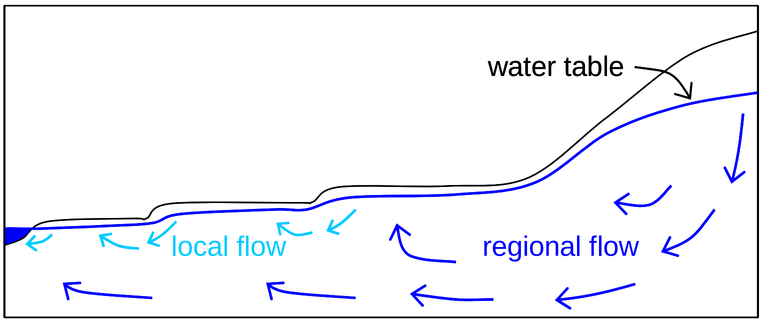

Figure 2.9. In coastal terrain, small local groundwater flow systems associated with terraces overlie more regional groundwater flow systems (adapted from Winter et al., 1998).

2.5.4. Glacial and dune terrain

Glacial and dune terrain is characterized by a landscape of hills and depressions. Although stream networks drain parts of these landscapes, many areas of glacial and dune terrain do not contribute runoff to an integrated surface drainage network. Instead, surface runoff from precipitation falling on the landscape accumulates in the depressions, commonly resulting in the presence of lakes and wetlands. Because of the lack of stream outlets, the water balance of these 'closed' types of lakes and wetlands is controlled largely by exchange of water with the atmosphere (precipitation and evapotranspiration) and with ground water.

Lakes and wetlands in glacial and dune terrain can have inflow from ground water, outflow to ground water, or both. The interaction between lakes and wetlands and groundwater is determined to a large extent by their position with respect to local and regional groundwater flow systems. A common conception is that lakes and wetlands that are present in topographically high areas recharge groundwater, and that lakes and wetlands that are present in low areas receive discharge from groundwater. However, lakes and wetlands underlain by deposits having low permeability can receive discharge from local groundwater flow systems even if they are located in a regional groundwater recharge area. Conversely, they can lose water to local groundwater flow systems even if they are located in a regional groundwater discharge area (Figure 2.10).

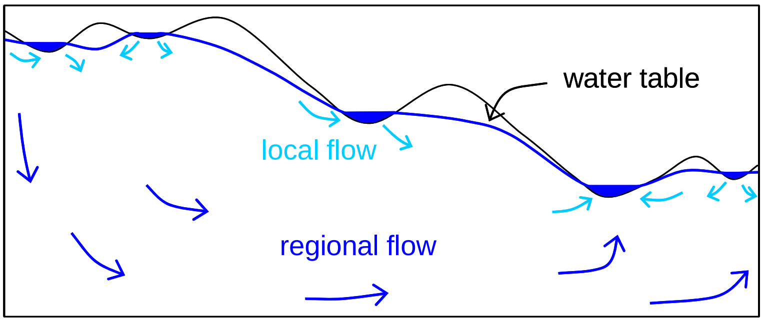

Figure 2.10. In glacial and dune terrain, local, intermediate, and regional groundwater flow systems interact with lakes and wetlands. It is not uncommon for wetlands that recharge local groundwater flow systems to be present in lowlands and for wetlands that receive discharge from local groundwater to be present in uplands (adapted from Winter et al., 1998).

2.5.5. Karst terrain

Karst may be broadly defined as all landforms that are produced primarily by the dissolution of rocks, mainly limestone and dolomite. Karst terrains are characterized by (1) closed surface depressions of various sizes and shapes known as sinkholes, (2) an underground drainage network that consists of solution openings that range in size from enlarged cracks in the rock to large caves, and (3) highly disrupted surface drainage systems, which relate directly to the unique character of the underground drainage system.

Groundwater recharge is very efficient in karst terrain because precipitation readily infiltrates through the rock openings that intersect the land surface. Water moves at greatly different rates through karst aquifers. It moves slowly through fine fractures and pores and rapidly through solution-enlarged fractures and conduits. As a result, the water discharging from many springs in karst terrain may be a combination of relatively slow-moving water draining from pores and rapidly moving storm-derived water.

Water movement in karst terrain is especially unpredictable because of the many paths groundwater takes through the maze of fractures and solution openings in the rock. Because of the large size of interconnected openings in well-developed karst systems, karst terrain can have true underground streams. These underground streams can have high rates of flow, in some places as great as rates of flow in surface streams. Furthermore, it is not unusual for medium-sized streams to disappear into the rock openings, thereby completely disrupting the surface drainage system, and to reappear at the surface at another place. Seeps and springs of all sizes are characteristic features of karst terrains. Springs having sufficiently large groundwater recharge areas commonly are the source of small- to medium-sized streams and constitute a large part of tributary flow to larger streams. In addition, the location where the streams emerge can change, depending on the spatial distribution of groundwater recharge in relation to individual precipitation events. Large spring inflows to streams in karst terrain contrast sharply with the generally more diffuse groundwater inflow characteristic of streams flowing across sand and gravel aquifers.

3. Groundwater flow models

Groundwater models provide a means to describe and simulate the complex patterns of subsurface water flow in the various settings detailed in the previous section. Groundwater models are conceptual descriptions or approximations of the physical system that employ mathematical equations based on certain simplifying assumptions to represent the system of interest. Assumptions usually involve the direction of flow (1-dimensional, 2-D or 3-D flow), the geometry of the aquifer, and the heterogeneity and anisotropy of the sediments or bedrock. Even the most elaborate spatially-distributed physically-based groundwater model is unable to exactly describe the true physics of the system. As a consequence, the model user must always understand the implications of the underlying assumptions of the model applied.

3.1. Groundwater flow equations

The first step in the modelling exercise is to build the conceptual model based on the perceptual understanding of the system. Next, the conceptual model is translated into the mathematical equations that comprise the mathematical model. A general form of the groundwater flow equation describing 3-D transient flow, derived by combining Darcy's law and the law of fluid mass conservation, is given by

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) = S_s \frac{\partial h}{\partial t} - R \tag{3.1}$$

where K_x, K_y and K_z are the components of the hydraulic conductivity tensor, h is the hydraulic head, S_s is the specific storage, or the volume of water that is released from storage in a unit volume of aquifer per unit decline in head, and R is a general fluid sink/source term (e.g., recharge, pumping). Based on the peculiarities of the system under study different forms of the above equation may apply.

3.2. Solution techniques

The governing groundwater flow equations are usually solved either analytically or numerically. Analytical methods provide exact solutions to the mathematical problem that are continuous in space and time, but they are limited in their applicability due to the stringent underlying assumptions required to implement them. In numerical models a discrete solution is obtained in space and time by employing numerical approximations of the governing partial differential equations. Numerical models are much more versatile and, with the unabated increase in computer power and the improvement of solution techniques, have outclassed all other types of groundwater models. Various numerical solution techniques are used in groundwater modelling. The most extensively used techniques are the method of finite differences, the method of finite elements and the analytical elements method.

3.3. Model codes

The set of commands used to solve the mathematical flow model on a computer forms the computer program or code. The code is generic, and it is transformed into a groundwater model by incorporating the site-specific geometry and boundary conditions and introducing the actual flow parameters. There is a plethora of groundwater computer codes available. They vary in the solution techniques used (e.g., analytical vs numerical; finite element vs finite difference methods), the dimension in which the governing equations are solved (1-, 2- and 3D), the iterative procedures used to solve the differential equations (e.g., strongly implicit procedure, slice successive over-relaxation), single or variable density flow, etc. All techniques have their own advantages and disadvantages with respect to availability, costs, user friendliness, applicability, and required knowledge of the user. It is not the intention to provide a complete overview of the list of existing groundwater modelling codes. The remainder of this section only elaborates on MODFLOW, the groundwater code that is most widely used and that is generally recognized as the standard code for groundwater modelling practices.

MODFLOW is a computer program originally developed by the U.S. Geological Survey that simulates three-dimensional groundwater flow using a finite difference technique for solution of the governing flow equations [Harbaugh et al., 2000]. MODFLOW solves both confined and unconfined flow equations in an irregularly shaped flow system to simulate the behaviour of groundwater flow systems under several types of natural and artificial stresses. The flow region is subdivided into blocks in which the medium properties are assumed to be uniform. In plan view the blocks are made from a grid of mutually perpendicular lines that may be variably spaced. Model layers can have varying thickness. A flow equation is written for each block, called a cell. Several solvers are provided for solving the resulting matrix problem. The user can choose the best solver for the particular problem. Flow-rate and cumulative-volume balances from each type of inflow and outflow are computed for each time step. Flow from external stresses, such as flow to wells, areal recharge, evapotranspiration, flow to drains, and flow through riverbeds, can be simulated. Hydraulic conductivities or transmissivities for any layer may differ spatially and be anisotropic, and the storage coefficient may be heterogeneous. Specified head, specified flux, and head dependent flux boundaries can be simulated.

Advantages of MODFLOW include numerous facilities for data preparation, the modular structure that allows it to be easily modified to adapt the code for a particular application, great flexibility in handling a wide range of complexity, easy exchange of data in standard form, extended worldwide experience, continuous development and availability of the source code. In addition to simulating groundwater flow, the scope of MODFLOW has been expanded to incorporate related capabilities such as solute transport and parameter estimation.

4. Overview of the representation of groundwater in regional scale catchment modelling

In regional scale catchment modelling the representation of the groundwater component ranges from simple conceptual (linear reservoirs) to complex physically-based (solving flow equations in 3 dimensions with complex boundary conditions) models and from lumped (aggregating spatial and temporal processes) to fully distributed models. Below, an overview is presented of the literature related to the representation of the groundwater component in regional scale catchment modelling. This overview is not intended to be exhaustive, but only mentions the most relevant works. For a complete overview on mathematical models of watershed hydrology the reader is referred to the excellent books edited by Singh [1995] and Singh and Frevert [2002a, 2002b].

4.1. Physically-based methods

Physically-based models solve a form of the groundwater flow equation (equation 3.1), given boundary and initial conditions. They require a full description of the aquifer system based on geological information, and the specification of hydraulic parameters (hydraulic conductivity, porosity, storage parameters) for each block or element in the

area. They yield a complete distribution of the groundwater head (in one, two or three dimensions) from which the direction of groundwater flow can be readily obtained.
Although MODFLOW is mainly devised to simulate groundwater flow, simplified river channels can be simulated using the MODFLOW stream package. The MODFLOW stream package is not actually a surface water package but does calculate channel-aquifer interaction and channel stage using the Manning's equation. Overland flow and flow in the unsaturated zone cannot be simulated. As such, MODFLOW as a stand-alone code is not capable of simulating surface water groundwater interaction at the catchment scale. However, a number of successful attempts that link surface water or watershed models with MODFLOW have appeared in the literature, e.g., SWATMOD [Sophocleous et al., 1999].

SWATMOD links the widely used watershed-scale model SWAT with MODFLOW. The linked models are used to simulate long-term surface water and groundwater interactions. SWATMOD was originally developed as an integrated surface water/groundwater model that could model an aquifer with distributed parameters and variable pumping. SWATMOD is a physically-based model operating on a watershed scale and capable of long time-period simulations. A limitation in the model design is the inability to model the unsaturated zone beyond the root zone. Therefore percolation (recharge) is applied directly to the groundwater table. Sophocleous et al. [1999] applied the coupled model for basin-wide management of a 3625-km^2 basin in Kansas.

Probably the most elaborate physically-based, spatially-distributed integrated surface water groundwater model is MIKE SHE of the Danish Hydraulic Institute (DHI). With additional DHI programs (MIKE 11 and MOUSE) that are easily linked to MIKE SHE, the capabilities of MIKE SHE are further expanded. MIKE SHE is used to simulate flow and transport of solutes and sediments in both surface water and groundwater. The water movement component of MIKE SHE comprises of five modules: evapotranspiration (ET), unsaturated zone flow (UZ), saturated zone flow (SZ), overland and channel flow (OC), and irrigation (IR). Several additional add-on modules are available for particle tracking, contaminant transport, soil plant systems, and other specialized modelling applications [DHI, 1999]. MIKE SHE includes a traditional 2D or 3D finite-difference groundwater model, which is very similar to MODFLOW.

Available boundary conditions are also comparable to those available in MODFLOW (i.e., wells, drains, etc.).

MIKE SHE has been successfully applied to small scale problems [e.g., Thompson et al., 2004] and to simulate the land-phase of the hydrological cycle for small [Singh et al., 1999], medium [Feyen et al., 2000] to large scale watersheds, and at the regional, or multi-watershed, scale [Henriksen et al., 2003].

The application of groundwater models that solve the governing groundwater flow equations in 2- or 3D, especially at large scales, is hampered by the vast amount of data required to represent the spatial variability of the hydrogeological characteristics and because of limits in computation recourses. Details about the sequence, lithology, thickness and structure of the rock formations that determine the occurrence, storage and movement of groundwater are usually lacking. Also, for certain purposes, e.g., to simulate the discharges at the catchment outlet or internal stations, it may not be needed to simulate hydrogeological processes based on physics or in great spatial detail, and lumped conceptual approaches may yield comparable results.

4.2. Simplified groundwater module of the Soil Water Assessment Tool

In the Soil and Water Assessment Tool (SWAT) [Neitsch et al., 2002] the groundwater component is represented using a two-aquifer system. The shallow aquifer is an unconfined aquifer that contributes to flow in the main channel or reach of the subbasin. The deep aquifer is a confined aquifer. Water that enters the deep aquifer is assumed to contribute to streamflow somewhere outside of the watershed [Arnold et al., 1993]. For the unconfined aquifer a linear storage equation is used to predict the non-steady-state response of groundwater flow to periodic recharge, assuming that the variation in return flow is linearly related to the rate of change in water table height (only headlosses in the horizontal direction are considered). The change in groundwater return flow is proportional to the excess recharge, with the proportionality or reaction factor being a function of the transmissivity of the aquifer and the slope length. As such, the reaction factor has a physical meaning, and can in principle be determined from field measurements. However, measurements of the hydraulic conductivity and the specific yield at the grid-scales used to model meso- to macro-scale basins cannot be readily obtained. This can be circumvented by determining the reaction factor from discharge measurements [Arnold et al., 1993] or observations of the fluctuation of the water table elevation [Hattermann et al., 2004]. This simplified procedure allows reproduction of the daily groundwater dynamics (water level and discharge) on a meso-scale and can be parameterized using meaningful data. More details on this approach are presented in Section 6.

4.3. Reservoir or storage models

Simpler conceptual models representing the catchment as a combination of interconnected stores and fluxes often describe the groundwater component as a single linear reservoir. The outflow of the reservoir is then a linear function of the storage in the reservoir, characterized by a storage constant (average residence time). In some cases baseflow is considered to be generated by a reservoir with multiple outlets or by a series of two or more storage reservoirs. This may result in a non-linear relation between groundwater storage and discharge. Examples of linear storage representations are the HBV-96 model [Lindström, 1997], the Sacramento Soil Moisture Accounting model [Burnash et al., 1973], the Tank model [Sugawara, 1995], and the LISFLOOD model [De Roo et al., 2000]. These types of models do not allow simulating the spatial distribution of hydraulic head, hence a comparison with observed data is difficult.

5. Current representation of groundwater in LISFLOOD

In the current version of LISFLOOD the groundwater part of the hydrological cycle is modelled employing two quasi-linear reservoirs, as shown in Figure 5.1. The upper groundwater zone represents the quick groundwater component to runoff. It is recharged by percolation from the subsoil ($D_{ls,ugw}$) and preferential flow through cracks and macro-pores in the soil ($D_{pref,gw}$). The lower groundwater zone represents the slow or delayed groundwater response that generates base flow. It is fed by drainage from the upper groundwater zone ($D_{ugw,lgw}$).

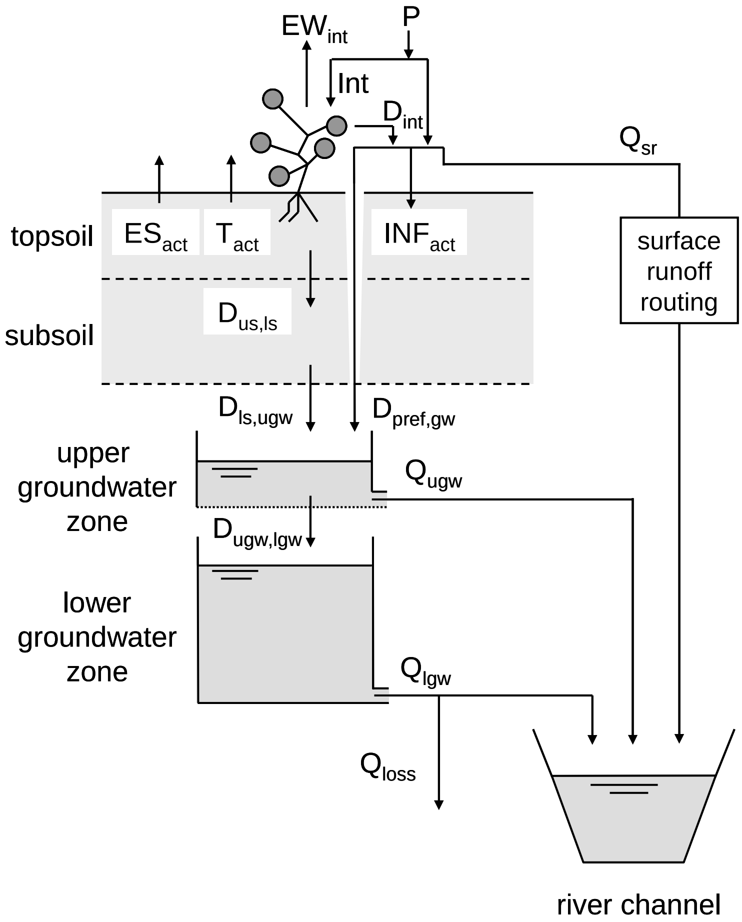

Figure 5.1. Schematic overview of the LISFLOOD model. P = precipitation; Int = interception; EW_{int} = evaporation of intercepted water; D_{int} = leaf drainage; ES_{act} = evaporation from soil surface; T_{act} = transpiration (water uptake by plant roots); INF_{act} = infiltration; Q_{rs} = surface runoff; $D_{us,ls}$ = drainage from upper to lower soil zone; $D_{ls,ugw}$ = drainage from lower soil zone to upper groundwater zone; $D_{pref,gw}$ = preferential flow to upper groundwater zone; $D_{ugw,lgw}$ = drainage from upper to lower groundwater zone; Q_{ugw} = outflow from upper groundwater zone; Q_{lgw} = outflow from lower groundwater zone; Q_{loss} = loss from lower groundwater zone. Note that snowmelt is not included in the Figure, even though it is simulated by the model).

Outflow from the upper groundwater zone, Q_{ugw} [mm], is given by

$$Q_{ugw} = \frac{1}{T_{ugw}} . S_{ugw}{}^{(1+a)} . \Delta t \, , \qquad (5.1)$$

were T_{ugw} is the upper zone reservoir constant [day.mm^a], S_{ugw} is the amount of water stored in the upper zone [mm], and a is a parameter that controls the degree of non-linearity between the outflow and storage of the reservoir. For a > 0, outflow calculated using equation (5.1) may exceed the storage. Also, the dimension of T_{ugw} depends on a. Therefore, the non-linearity parameter is always set to one, for which equation (5.1) reduces to a linear relation between the reservoir outflow and storage.

The upper reservoir also recharges the lower reservoir. For each time step, a fixed amount of water percolates from the upper to the lower zone

$$D_{ugw,lgw} = \min(GW_{perc}.\Delta t, S_{ugw}) \tag{5.2}$$

were GW_{perc} [mm.day^{-1}] is the maximum amount of water entering the lower zone within a time step.

Outflow from the lower groundwater zone, Q_{lgw} [mm], is given by the following linear relation

$$Q_{lgw} = \frac{1}{T_{lgw}}.S_{lgw}.\Delta t\,, \tag{5.3}$$

where T_{lgw} is the lower zone reservoir constant [day.mm], and S_{lgw} is the amount of water stored in the lower zone [mm].

For each time step, storage in the upper and lower zone is updated for the incoming and outgoing fluxes using the following water balance equations

$$S_{ugw,t} = S_{ugw,t-1} + D_{pref,gw} - D_{ugw,lgw} - Q_{ugw} \tag{5.4}$$

$$S_{lgw,t} = S_{lgw,t-1} + D_{ugw,lgw} - Q_{lgw} \tag{5.5}$$

Within each time step the outflow from the upper (Q_{ugw}) and lower (Q_{lgw}) groundwater zone is routed to the nearest downstream channel pixel. The routing procedure is illustrated in Figure 5.2. For each pixel that contains a river channel, the drainage network defines the contributing pixels. The total inflow of groundwater to a river channel pixel then equals the sum of the outflows of the individual upstream pixels. For example, two flow paths contribute to the river channel node in the second left column of the bottom row in Figure 5.2. Hence, the inflow to this river pixel within a certain time step equals the sum of the outflows from the upper and lower zone of all pixels along the flow paths, which are denoted by q_1 and q_2, respectively.

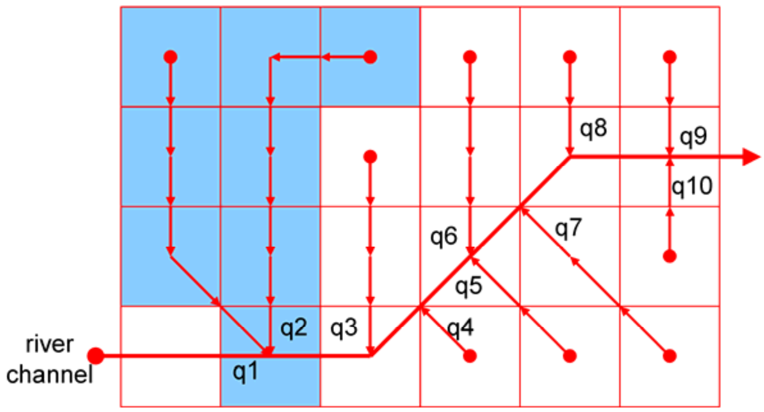

Figure 5.2. Schematic overview of the routing process of groundwater flow in LISFLOOD.

Table 5.1 presents an overview of the parameters that need to be specified by the user for the current groundwater component of LISFLOOD. The values of T_{ugw}, T_{lgw} and GW_{perc} lack physical basis, hence they can not be obtained from measurements and should be determined by calibration.

Table 5.1. Overview of parameters that need to be specified for the current representation of the groundwater component in LISFLOOD.

Variable	Definition
T_{ugw}	Upper zone reservoir constant
T_{lgw}	Lower zone reservoir constant
GW_{perc}	Maximum amount of water percolating from upper to lower zone

6. Proposal to improve groundwater representation in LISFLOOD

The groundwater component of LISFLOOD must have the level of sophistication similar to those of the other components represented in the model. Therefore, a detailed numerical model is not justified in this case, and a relatively simple yet realistic approach should be adapted. A complex groundwater model at the regional or European scale is also infeasible due to the lack of data, the geological complexity across Europe, the enormous work this would require (e.g., interpreting geological profiles, deriving flow and storage properties), and computational limitations.

In the current LISFLOOD setup, outflow from the upper groundwater zone more likely reflects interflow rather than groundwater flow, hence does not really contribute to the slow response of the hydrograph that represents groundwater baseflow. This may have some implications for adapting the groundwater component of the LISFLOOD model.

In the remainder of this section, two methods are proposed. The first method is based on the current groundwater representation in LISFLOOD using the two interconnected reservoirs, using a simple conversion to translate the storage in the lower groundwater reservoir into estimates of the watertable elevation. The second method is based on the two layer aquifer system used in the Soil and Water Assessment Tool (SWAT).

6.1. Converting lower groundwater storage into water table elevations

As mentioned above, outflow from the upper groundwater reservoir can be considered as interflow. Therefore, storages in this reservoir can not be considered as being part of the groundwater storage. Therefore, we only consider the lower groundwater zone. Outflow from the lower groundwater zone, Q_{lgw} [mm], is given by the following linear relation

$$Q_{lgw} = \frac{1}{T_{lgw}}.S_{lgw}.\Delta t \,, \tag{6.1}$$

where T_{lgw} is the lower zone reservoir constant [day.mm], and S_{lgw} is the amount of water stored in the lower zone [mm]. For each time step, storage in the lower zone is updated for the incoming and outgoing fluxes using the following water balance equation

$$S_{lgw,t} = S_{lgw,t-1} + D_{ugw,lgw} - Q_{lgw} \tag{6.2}$$

Since the volume of groundwater storage is reflected by the height of the water table, variations in storage are reflected by variations in groundwater level. The water stored in a unit volume aquifer equals the product of the water level height with the porosity (n), hence modelled storages could be converted into water levels through division by the porosity. However, some of the pores spaces may be too small or too poorly connected to permit the water they contain to flow out easily. Therefore, effective porosity (n_e) should be used to convert the storages into water levels. The effective porosity can be thought of as the volume of pore space that will drain under the influence of gravity.

This procedure yields an estimate of the water table elevation. However, the estimated hydraulic head elevations will only reflect the true (or observed) elevations when the storage in the lower groundwater reservoir reflects the true storage in the aquifer. This implies that the bottom of the storage reservoir coincides with the true base of the aquifer. Given the complexity of groundwater aquifer systems, and the sparse information typically available, it is not convenient to derive the level of an aquifer base, especially at the regional or European scale. This problem can be circumvented by assuming that the initial water levels (converted from the initial storage) in the lower reservoir at a certain step in time (e.g., October 1st), after a sufficiently long initialisation and warming up period, corresponds to the water level observations for that day (October 1st). Simulated variations of the groundwater level should then be added to the observed values at the start of the actual simulation (October 1st).

The single reservoir model may be a crude approximation for complex aquifer systems with varying porosity over depth and a more non-linear response to storage and groundwater level. In this case reference levels for a stepwise increase in runoff can be specified, with a different effective porosity for each zone.

Table 6.1 gives an overview of the parameters that need to be specified for the updated groundwater component. This representation would only introduce one additional parameter, namely the effective porosity of the aquifer to convert reservoir storage into hydraulic head elevation. It must be stressed that the values of the effective porosity are model parameters, that may not necessarily coincide with the true effective porosity of the system, as they may be very well affected by shortcomings in the storage reservoir representation (or other model structure errors) of the groundwater system. The estimation of spatially distributed water levels allows observations of the water table elevation to be used in the calibration of the model, not only to determine the effective porosity, but potentially also to improve the estimates of the lower zone reservoir constant.

Table 6.1. Overview of parameters that need to be specified for the proposed representation of the groundwater component in LISFLOOD.

Variable	Definition
T_{ugw}	Upper zone reservoir constant
T_{lgw}	Lower zone reservoir constant
n_e	Effective porosity
GW_{perc}	Maximum amount of water percolating from upper to lower zone

6.2. Two aquifer system of SWAT

The second methodology proposed to potentially improve the representation of the groundwater component in LISFLOOD is based on the two-aquifer system of the Soil and Water Assessment Tool (SWAT). SWAT is a river basin scale model developed to quantify the impact of land management practices in large, complex watersheds. It is a public domain model actively supported by the USDA Agricultural Research Service at the Grassland, Soil and Water Research Laboratory in Temple, Texas, USA [Neitsch et al., 2002].

In the method proposed the groundwater component is modelled as a system of two aquifers for each subbasin. The shallow aquifer is an unconfined aquifer that contributes to flow in the main channel or reach of the subbasin. The deep aquifer is a confined aquifer. Water that enters the deep aquifer is assumed to contribute to streamflow somewhere outside of the watershed [Arnold et al., 1993].

6.2.1. Shallow aquifer

The water balance equation for the shallow aquifer is given by

$$S_{sa,t} = S_{sa,t-1} + D_{sa} - D_{sa,da} - D_{revap} - Q_{sa,pump} - Q_{sa} \tag{6.3}$$

where $S_{sa,t}$ is the amount of water stored in the shallow aquifer at time t [mm], $S_{sa,t-1}$ is the amount of water stored in the shallow aquifer at time t-1 [mm], D_{sa} is the amount of water recharging the shallow aquifer within time step t [mm], $D_{sa,da}$ is the amount of water that percolates to the deep aquifer within time step t [mm], D_{revap} is the amount of water that moves back up into the soil zone within time step t and that is lost to the atmosphere by soil evaporation or plant root uptake [mm], $Q_{sa,pump}$ is the amount of water extracted by pumping from the shallow aquifer within time step t [mm], and Q_{sa} is the groundwater contribution to stream flow, or base flow, within time step t [mm]. In what follows, the different components of the water balance for the shallow aquifer are detailed.

6.2.2. Recharge to the shallow aquifer

In SWAT, the shallow aquifer is recharged by water that percolates from the soil zone (D_{sa}). However, recharge may not always occur instantaneously (i.e., within one time step). Depending on the depth to the water table and the hydraulic properties of the system there may be a lag between the moment that water exits the soil profile and when it actually reaches the water table. The lag time is accommodated for by using the following exponential delay function [Venetis, 1969; Sangrey et al., 1984]

$$D_{sa,t} = \left(1 - \exp\left[{-1}/{\delta_{gw}}\right]\right) \cdot D_{seep} + \exp\left[{-1}/{\delta_{gw}}\right] \cdot D_{sa,t-1} \tag{6.4}$$

where $D_{sa,t}$ is the amount of recharge entering the shallow aquifer at time t [mm], $D_{sa,t-1}$ is the amount of recharge entering the shallow aquifer at time t-1 [mm], D_{seep} is the outflow from the soil zone within time step t [mm], and δ_{gw} is the delay time or drainage time of the unsaturated zone between the soil zone and groundwater table. The latter cannot be measured directly, hence it needs to be determined by calibration. This can be done by comparison of simulated variations in the water table level with observations. It is important to note that equation 6.4 affects only the timing of groundwater recharge, and therefore return flow, and not the total volume.

6.2.3. Groundwater contribution to stream discharge

The shallow aquifer contributes to the stream discharge through base flow when the amount of water stored in the shallow aquifer exceeds a user-specified threshold value $S_{sa,thr}^{bf}$. The approach proposed to estimate base flow from the shallow aquifer is that of Smedema and Rycroft [1983]. They derived a linear storage equation to predict the non-steady response of groundwater flow to periodic recharge based on Hooghoudt's [1943] formula of steady-state response of groundwater flow to recharge

$$Q_{sa} = \frac{8K_s}{L^2} \cdot h \tag{6.5}$$

where K_s is the hydraulic conductivity of the shallow aquifer [m/day], h is the water table height [m], and L is the slope length [m]. Assuming that the variation in return flow Q_{sa} per time step is linearly related to the rate of change in water table height h (only headlosses in horizontal direction are considered), this gives

$$\frac{dQ_{sa}}{dt} = \frac{8K_s}{L^2} \cdot \frac{dh}{dt} \tag{6.6}$$

When the groundwater body is recharged (D_{sa}), either by soil percolation or bypass flow, and depleted by drain discharge (Q_{sa}), it follows that the water table will rise when $D_{sa} - Q_{sa} > 0$ and fall when $D_{sa} - Q_{sa} < 0$. Fluctuations of the water table are given by [Smedema and Rycroft, 1983]

$$\frac{dh}{dt} = \frac{(D_{sa} - Q_{sa})}{0.8 \cdot S_y} \tag{6.7}$$

where S_y is the specific yield of the shallow aquifer. From 6.6 and 6.7, it follows that

$$\frac{dQ_{sa}}{dt} = \frac{10K_s}{S_y \cdot L^2} \cdot (D_{sa} - Q_{sa}) = \alpha \cdot (D_{sa} - Q_{sa}), \tag{6.8}$$

or that the change in groundwater return flow is proportional to the excess recharge, with α being the proportionality factor or reaction factor. After separation of the variables, integration of equation 6.8 between time steps t and t-1 gives

$$\frac{D_{sa} - Q_{sa,t}}{D_{sa} - Q_{sa,t-1}} = \exp(-\alpha \cdot \Delta t) \tag{6.9}$$

Rearranging equation 6.9 yields an expression for the return flow at time t

$$Q_{sa,t} = Q_{sa,t-1} \cdot \exp(-\alpha \cdot \Delta t) + D_{sa}(1 - \exp(-\alpha \cdot \Delta t)) \tag{6.10}$$

Using the linear relationship between Q_{sa} and h we get the expression for the water table level

$$h_t = h_{t-1} \cdot \exp(-\alpha \cdot \Delta t) + \frac{D_{sa}}{0.8 \cdot S_y \cdot \alpha}(1 - \exp(-\alpha \cdot \Delta t)) \tag{6.11}$$

The reaction factor or baseflow recession constant α is a direct index of the groundwater flow response to changes in recharge [Smedema and Rycroft, 1983]. Typical values range from 0.1-0.3 for areas with a slow response to recharge to 0.9-1.0 for areas with a quick response. The reaction factor is a function of the hydraulic conductivity and specific yield of the shallow aquifer and of the slope length

$$\alpha = \frac{10K_s}{S_y \cdot L^2} \tag{6.12}$$

Therefore, the reaction factor α has a physical basis and in principle can be calculated based on field measurements. However, measurements of the hydraulic conductivity and the specific yield at the grid-scales used to model meso- to macro-scale basins cannot be readily obtained. This can be circumvented by determining α from discharge measurements [Arnold et al., 1993] or observations of the fluctuation of the water table elevation [Hattermann et al., 2004]. In periods with no recharge equation 6.10 reduces to

$$Q_{sa,t} = Q_{sa,t-1} \cdot \exp(-\alpha \cdot \Delta t) \tag{6.13}$$

from which the following expression for the baseflow or recession constant is obtained

$$\alpha = \frac{\ln Q_{sa,t-1} - \ln Q_{sa,t}}{\Delta t} \tag{6.14}$$

The recession constant may thus be estimated by analysing the stream discharge during periods of no recharge ($D_{sa} = 0$).

Hattermann et al. [2004] suggested an alternative method using the linear relationship between Q_{sa} and h, for which equation 6.14 becomes

$$\alpha = \frac{\ln h_{t-1} - \ln h_t}{\Delta t} \tag{6.15}$$

Based on equation 6.15, α can be estimated using observations of the water table elevation. Observations of the water table are usually more abundant and spatially distributed than stream discharge observations. Hence, the calibration of the model against water table measurements will likely improve the predictions of stream discharge.

6.2.4. Upward movement of water back into the soil zone

Water is allowed to move from the shallow aquifer back to the overlying soil zone. In dry periods, water in the capillary fringe between the saturated and unsaturated zone moves upward and is replaced by water from the shallow aquifer. Water may also be removed from the shallow aquifer by direct plant-root uptake. These processes are significant in aquifer systems where the water table is close to the ground surface or where plants with deep roots are prevalent. The movement of water to the soil profile is termed 'revap' and is modelled as a function of the water demand for evapotranspiration. Revap is only allowed to occur when the amount of water stored in the shallow aquifer exceeds a user-specified threshold value $S_{sa,thr}$. The maximum amount of water that can be extracted from the shallow aquifer by revap within a time step is given by

$$D_{revap,\max} = \beta_{rev} \cdot ET_0 \tag{6.16}$$

where $D_{revap,\max}$ is the maximum amount of water that moves up due to water deficiencies in the soil zone [mm], β_{rev} is the revap coefficient, and ET_0 is the potential evapotranspiration [mm]. The actual amount of water that moves up through revap within a time step is given by

$$D_{revap} = 0 \qquad \text{if } S_{sa} < S_{sa,thr} \tag{6.17a}$$

$$D_{revap} = D_{revap,\max} - S_{sa,thr} \qquad \text{if } S_{sa,thr} < S_{sa} < (S_{sa,thr} + D_{revap,\max}) \tag{6.17b}$$

$$D_{revap} = D_{revap,\max} \qquad \text{if } S_{sa} \geq (S_{sa,thr} + D_{revap,\max}) \tag{6.17c}$$

6.2.5. Percolation of groundwater from the shallow to the deep aquifer

A portion of the water stored in shallow aquifers recharges deep aquifers. Percolation to the deep aquifer is allowed only when the amount of water stored in the shallow aquifer exceeds the user-specified threshold value $S_{sa,thr}$. The maximum amount of water that can percolate to the deep aquifer within a time step is given by

$$D_{sa,da,\max} = \beta_{da} \cdot D_{sa} \tag{6.18}$$

where $D_{sa,da,\max}$ is the maximum amount of water that percolates to the deep aquifer [mm], and β_{da} is the deep aquifer percolation coefficient. The actual amount of water that is lost from the shallow aquifer by percolation to the deep aquifer within a time step is given by

$$D_{sa,da} = 0 \qquad \text{if } S_{sa} < S_{sa,thr} \tag{6.19a}$$

$$D_{sa,da} = D_{sa,da,\max} - S_{sa,thr} \qquad \text{if } S_{sa,thr} < S_{sa} < (S_{sa,thr} + D_{sa,da,\max}) \tag{6.19b}$$

$$D_{sa,da} = D_{sa,da,\max} \qquad \text{if } S_{sa} \geq (S_{sa,thr} + D_{sa,da,\max}) \tag{6.19c}$$

6.2.6. Groundwater extraction from the shallow aquifer by pumping

Because water management affects the hydrologic balance, it is critical that the model is able to accommodate for management practices. For example, in some areas extensive pumping from the shallow aquifer takes place for irrigation or drinking water purposes, Water removed for consumptive use can be considered to be lost from the system. The term $Q_{sa,pump}$ in the water balance equation 6.3 allows accounting for groundwater extraction from the shallow aquifer through pumping. The value specified for $Q_{sa,pump}$ should be based on information about groundwater extraction rates. An amount up to the total volume of water stored in the shallow aquifer can be extracted within a time step.

6.2.7. Deep aquifer

For the deep aquifer, the water balance equation is given by

$$S_{da,t} = S_{da,t-1} + D_{sa,da} - Q_{da,pump} \tag{6.20}$$

where $S_{da,t}$ is the amount of water stored in the deep aquifer at time t [mm], $S_{da,t-1}$ is the amount of storage in the deep aquifer at time t-1 [mm], and $Q_{da,pump}$ is the amount of water extracted from the deep aquifer by pumping. The amount of water percolating to the deep aquifer is calculated with equation 6.19. An amount up to the total volume of water stored in the deep aquifer can be extracted within a time step. Water that enters the deep aquifer is assumed to be lost from the system, hence is not considered in future water budgets.

Table 6.2 gives an overview of the parameters that need to be specified for the 2 layer aquifer system. The specific yield and recession constant can be obtained from field measurements if available, but most likely all parameters need to be determined by calibration against discharge and water level observations.

Table 6.2. Overview of parameters that need to be specified for the proposed representation of the groundwater component in LISFLOOD

Variable	Definition
δ_{gw}	Delay time for shallow aquifer recharge
$S_{sa,thr}^{bf}$	Threshold water level in shallow aquifer for base flow
β_{rev}	Revap coefficient
β_{da}	Deep aquifer percolation coefficient
$S_{sa,thr}$	Threshold level in shallow aquifer for revap and percolation to deep aquifer
S_Y	Specific yield of the unconfined aquifer
α	Base flow recession constant

Implementation of the above approach is not as straightforward as the first method proposed. In SWAT, the shallow aquifer is recharged (D_{sa}) by water that percolates from the soil and water that bypasses the soil profile through preferential flow. Interflow is incorporated in the soil component of the SWAT model. Therefore, it is probably better to replace only the lower groundwater zone by the 2-layer aquifer model and to keep the upper groundwater zone in LISFLOOD to account for interflow. The shallow aquifer is then fed by outflow from the upper groundwater zone.

Another problem that requires attention in the implementation of this approach is the upward movement of water back into the soil profile or to the atmosphere (revap). Since the upper groundwater zone only accounts for interflow, revap should bypass this zone. Some other remaining issues that should be looked at carefully are:

- Should the upward movement of water bypass the soil component, i.e., be directly considered as lost to the atmosphere due to evapotranspiration?
- In LISFLOOD water for evapotranspiration can only be extracted from the upper soil layer, which implies that any upward movement from the groundwater to the lower soil component is not effective. Should evapotranspiration be allowed from the second soil zone, or should the soil be represented by one single layer?
- How to update soil storage for the upward flux?

7. Summary, conclusions and way forward

This document presented possible ways to adapt LISFLOOD to better represent the groundwater component in the model and to be able to simulate the distribution of groundwater levels in space and time. Detailed physically-based approaches, such as MODFLOW or the groundwater component used in MIKE SHE, require an enormous amount of data to represent the complex aquifer structure across Europe and to parameterize the model. The methods are also computationally very expensive, which hampers their application to regional or the European scale. Also, the groundwater component of LISFLOOD should have the level of sophistication similar to those of the other components represented in the model. Therefore, a detailed numerical model is not justified in this case, and a relatively simple yet realistic approach should be adapted.

Two approaches have been proposed. The first method is based on the current groundwater representation in LISFLOOD using the two interconnected reservoirs, using a simple conversion based on the effective porosity to translate the storage in the lower groundwater reservoir into estimates of the water table elevation. Only one additional parameter, the effective porosity, is introduced. For complex aquifer systems with varying porosity over depth and a more non-linear response to storage and groundwater level reference levels for a stepwise increase in runoff can be specified, with a different effective porosity for each zone.

In the second approach the lower groundwater reservoir is modelled as a system of two aquifers. The shallow aquifer is an unconfined aquifer that contributes to flow in the main channel or reach of the subbasin. The deep aquifer is a confined aquifer. Water that enters the deep aquifer is assumed to contribute to streamflow somewhere outside of the watershed. The change in groundwater return flow is proportional to the excess recharge, and is linearly related to the rate of change in water table height. The method also allows for a delay in the recharge to the unconfined aquifer, and a upward flux to the soil zone. The method is less compatible with the current setup of the LISFLOOD component, and 7 additional parameters are introduced. Some of those can be obtained from field data, but given the limited data on the regional and European scale, the likely need to be estimated by calibration. The advantage of both methods is that spatially distributed groundwater elevation observations can be used in the calibration process.

To test the validity of the methods, they should be tested on several catchments with varying complexity in aquifer structure and with a good spatial coverage of water table observations. Calibration and validation against water table elevations may also improve the understanding and shortcomings of the current model setup. If the approaches prove to be valuable, and allow effective estimation of the groundwater level fluctuations, the updated model can potentially be used in drought analysis and the assessment of climate and land use change impacts on the elevation of the groundwater table.

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