Review of groundwater–surfacewater interaction modelling approaches and their suitability for Australian conditions

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Executive summary

This report presents the outcomes of a review of the available approaches for modelling groundwater–surfacewater interactions at various spatial and temporal scales with different levels of complexity. It is aimed at determining the deficiencies of the current modelling approaches for Australian conditions and making recommendations for future model development. It is one of a series of three reports (the other two reports being the ‘Catalogue of conceptual models for groundwater–stream interaction’ (Reid et al. 2008) and ‘Estimation and prediction of the exchange of groundwater and surfacewater – field methodologies’ (Turner 2008)) that are the culmination of the first tasks of the CRC eWater D3 project ‘Estimation and prediction of the exchange of groundwater and surfacewater in lowland rivers’ and the associated National Water Commission Project ‘Australian Hydrological Modelling Initiative – Groundwater–SurfaceWater Interaction Tool (AHMI-GSWIT)’. These projects aim to predict exchange fluxes between groundwater and surfacewater for lowland rivers and predict how these may change with existing or different groundwater and surfacewater management. As such, these projects address a major recognised deficiency in the management of stressed or threatened Australian catchments in accounting for groundwater-stream interaction in the water budget.

The most important processes involved in groundwater–surfacewater interactions are first outlined. We then present studies cited in the literature that model such processes, and these studies are categorised according to model complexity. For this purpose, we define three levels of complexity, namely:

- Level 1. 1st order lumped parametric models based on empirical relations derived from numerous field observations or concepts.
- Level 2. 2nd order models that operate at finer temporal and spatial scales compared to 1st order models and have more conceptual resolution and process complexity.

We list the modelling packages available, identify the processes that they model, and rank them in terms of their complexity in a tabular form. We then provide an assessment of the available models and hence identify their deficiencies based on the suitability criteria of the D3/AHMI-GSWIT projects.

On the basis of these analyses, given the complexity of the processes of groundwater–surfacewater interaction, and the general scarcity of data required to validate and/or calibrate complex models, we recommend that a simple modelling approach be adopted for the D3/AHMI-GSWIT projects. We propose the development of:

1. A Level 1 complexity, reach scale, ‘GW–SW link’ model, which operates as a groundwater link to river models. The expected outcome of this model is accounting for GW–SW interactions at the river-reach scale.
2. A Level 2 complexity, sub-reach scale, ‘Floodplain Processes’ model, which dynamically models bank storage, evapotranspiration, and floodplain inundation. The expected outcome is modelling of GW–SW interactions at the sub-river-reach scale with higher resolution and the capacity to link to ecological response models.
1 Introduction

The D3/AHMI-GSWIT projects aim to predict exchange fluxes between groundwater and surfacewater for lowland rivers and how these may change with groundwater and surfacewater management. This capacity will be delivered as software models within the Catchment Modelling Toolkit (http://www.toolkit.net.au), which will be developed using The Invisible Modelling Environment (TIME) framework. Those models will mainly address issues related to the conjunctive management of groundwater and surfacewater and groundwater-dependent ecosystems. The eWater CRC has a strong mandate to develop and deliver software products using model development frameworks previously adopted by the CRC for Catchment Hydrology. The various modelling outputs from this project will be required to link with the eWater products that are highly relevant to the water industry in Australia. Of particular relevance are the products P2 (RiverManager Model/AHMI-River Systems Modelling Tool) and P5 (Water and Constituent Simulation Tool, WaterCAST). This document is targeted towards scientists and engineers involved in the assessment and management of water resources where groundwater–surfacewater interaction is a key component.

1.1 Objectives

There has been a large amount of work done overseas and within Australia on the processes affecting groundwater–surfacewater interactions. These processes can be complex and include, flooding recharge, evapotranspiration from shallow water tables, groundwater interception by wetlands, parafluvial flow, hyporheic exchange, bank storage effects caused by fluctuating river levels, groundwater extraction, structural features causing heterogeneity of flow. Methods have been developed to measure and model these processes at a range of spatial and temporal scales. However, it is unclear how relevant these methods are to Australia, which has unique landscape settings and an often sparse availability of groundwater and surfacewater data.

This document is a thorough literature review of processes related to groundwater–surfacewater (GW–SW) interactions and the available tools to model them at various spatial and temporal scales with different levels of complexity. It is intended to enhance awareness of the different modelling approaches published in the peer-reviewed literature, prevent duplication, and underpin adaptation and initiate informed debate. Because the D3/AHMI-GSWIT projects are concentrating on the issues related to integrated groundwater–surfacewater accounting and groundwater dependent ecosystems, the scope of the review is limited to water quantity and therefore does not address water quality aspects (i.e. salt, nutrients, pH, temperature, sediments etc.).

The main objective of this document is to review all of the available models, identify their suitability for Australian conditions, and therefore gauge the suitability of those models to the requirements of the D3/AHMI-GSWIT projects. Moreover, this document outlines the models that will be developed in these projects along with the data sets required for the application of these models. The default data sets should inform the design of field experimentation outlined in Turner (2007).

1.2 Model design criteria and suitability

The suitability of models to address the research questions posed in the D3/AHMI-GSWIT projects is assessed against a set of criteria that were deemed necessary for these modelling tools to be successfully applied to the relevant Australian landscape settings and at the correct spatial/temporal scales.

Reid et al. (2008) identified four landscape settings that are commonly encountered in Australia, namely: (1) upland fractured rock systems; (2) layered fractured rock systems e.g. sandstones, basalts; (3) contained alluvial valleys; (4) regional systems. It is imperative that the developed models are applicable to those landscape settings.
The spatial scale at which a model is applied dictates its level of complexity and hence what processes are, and aren’t accounted for. Large-scale models usually adopt a lumped approach that requires less parameterisation whereas smaller scale physically based models can explicitly account for more processes. Identifying the dimensionality of any problem is of vital importance as model complexity varies in an exponential manner with model dimensionality (a 10-element 1-D model, has 100 elements in a 2D model, and 1000 elements in a 3D model).

Data requirements are closely related to model complexity and the spatial scale at which the model operates. Lumped models require less data whereas process-based models require much more data. In many cases, model choice is restricted by data availability. At a whole-of-river scale, readily available data can support low fidelity modelling whereas intensive measurements at a sub-reach scale may be needed to support high fidelity modelling; this directly impacts field experimental design.

When choosing modelling tools for groundwater–surfacewater interaction, it is important to strike the right balance between surfacewater and groundwater processes. That is, we need to clearly define the problem and hence identify whether or not any emphasis should be placed on groundwater or surfacewater processes; this would clearly significantly impact model choice.

The issue of temporal scales becomes critical when modelling groundwater–surfacewater interaction as surfacewater processes are quick whereas groundwater processes are much more attenuated. Large time lumping in evaluating a particular process may mask other processes that may occur during short periods as a result of the averaging effect of larger time interval data. Large disparities in time steps between surfacewater and groundwater may lead to numerical instabilities.

Other issues that affect model suitability for the purposes of the D3/AHMI-GSWIT projects are whether or not software packages (and their source codes) are freely available or not. Another important criterion is the ability to develop the model within the TIME framework, which is the preferred option for the eWater CRC.

1.3 Approach

It is clear that the choice of a particular modelling tool will be largely impacted by its ability to model various processes relevant to groundwater–surfacewater interaction. It is also common knowledge that various software packages model different processes in varying levels of complexity.

Section 2 of this document outlines the most important processes involved in groundwater–surfacewater interactions. We present studies cited in the literature that model such processes; these studies are categorised according to model complexity. For this purpose, we define three levels of complexity, namely: (1) 1st order lumped parametric models based on empirical relations derived from numerous field observations or concepts, (2) 2nd order models that operate at finer temporal and spatial scales compared to 1st order models and have more conceptual resolution and process complexity, and (3) process-based, distributed, deterministic models. The varying levels of complexity relate closely to data availability. Simple lumped models require less data and usually operate at a large scale whereas process based models have high data requirements and are capable of making predictions at a much finer spatial and temporal scales.

In Section 3, we list the modelling packages available, identify the processes that they model, and rank them in terms of their complexity in a tabular form following the levels of complexity approach used in Section 2. Section 4 provides an assessment of the available models and hence identifies their deficiencies based on the criteria of the D3/AHMI-GSWIT projects defined in Section 1.2. Finally, Section 5 outlines the recommendations regarding the modelling activities that should be undertaken in these projects in order to successfully achieve their aspired goals.
2 Processes relevant to groundwater–surfacewater interactions, and modelling approaches used to represent them

In this section, we outline the most important processes related to groundwater–surfacewater interactions and the tools available model them. We categorise the processes as being groundwater driven (Section 2.1), surfacewater driven (Section 2.2), and groundwater–surfacewater driven (Section 2.3). Some of those processes are conceptually shown in Figure 1, which shows the interaction between a groundwater aquifer and a gaining stream. The tools for modelling each process are classed into three categories based on their level of complexity as outlined in Section 1.3. The principles of groundwater–surfacewater connectivity (i.e., hydraulic connection in gaining/losing streams) have been thoroughly discussed in Reid et al. (2008) and are hence not discussed in this document.

![Diagram of processes relevant to groundwater–surfacewater interaction](image)

Figure 1. Processes relevant to groundwater–surfacewater interaction.

### 2.1 Groundwater driven processes: stream depletion

Pumping-induced stream depletion is defined as the reduction of stream flow due to induced infiltration of stream water into the aquifer or the capture of aquifer discharge to the stream (Sophocleous 1997; Theis 1941). The problem of stream-aquifer interaction is pertinent to conjunctive-use management of water resources and riparian zone hydrology (Hantush 2005). This concept is only relevant to streams connected to the aquifer via a fully saturated material.

Pumping of groundwater near gaining streams has been shown to have considerable effects on water supply, water quality, and water rights administration. The environmental impacts of stream depletion are particularly important in small streams (Hunt 2003). During pumping, and while the cone of depression is progressing towards a nearby stream, groundwater depletion
occurs. When this cone reaches the stream, the rate of groundwater discharge to the stream reduces, and surfacewater may even start to infiltrate the aquifer; thus marking the start of stream depletion. After a long period of pumping, the cone of depression takes its final shape (i.e., steady state is reached), and a portion, or in some cases all, of the pumping will be balanced by a reduction in or reversal of flow from the aquifer to the stream. The proportion of pumping met by stream depletion in the steady state case will depend on various factors, including the proximity of the bore to the stream compared to the distance between the bore and other point and diffuse sources of groundwater recharge. Stream depletion in leaky aquifers is likely to only partially support groundwater withdrawal from a pumping well (Zlotnik 2004).

The time required to reach steady state varies with the square of the distance between the stream edge and the pumping well and varies linearly with aquifer diffusivity. Other important factors that may significantly affect stream depletion include streambed clogging as quantified by streambed-aquifer hydraulic conductivity contrast, degree of stream partial penetration, and aquifer heterogeneity. Identifying the time to reach steady state is important. For example, when pumping is for irrigation purposes, then if the time to reach steady state is a sizable portion of the irrigation season, then studying transient effects becomes critical to sound water management.

In theoretical terms, the water rights to withdraw aquifer water at early stages of pumping before the cone of depression reaches the stream become water rights to withdraw stream water at steady state (Sophocleous 2002). The analytical solution of Glover and Balmer (1954) is used to demonstrate the time scales for groundwater and surfacewater depletion during pumping. Figure 2 (a reproduction of Figure 9 from Sophocleous (2002)) demonstrates the concept of switching from reductions in groundwater storage at early times to stream depletion at later stages of pumping; note that complete reliance on surfacewater occurs only after extremely long times.

Figure 2. GW–SW depletion during pumping (using Glover and Balmer solution).

If surfacewater is polluted, then water quality becomes an important issue. In this case, we need to define the geometry of the ‘influence zone’ during pumping to offer information for wellhead protection. For example, when pumping is seasonal (e.g., for irrigation purposes), does the infiltrated stream water arrive at the pumping well or does it remain in the aquifer then drains back to the stream during the recovery period?

Figure 3 shows the flux response (representing stream depletion) at various times along the stream for a pumping well located 1,000 m from a stream (Rassam et al. 2005a); those flux response curves are analogous to the influence zones reported by Chen (2001) who obtained them by defining the path lines of the infiltrated stream water towards a pumping well. Chen (2001) reported that the existence of a regional hydraulic gradient towards a gaining stream
reduces the influence zone and that the influence zone during recovery is smaller than that during pumping when a gradient is present. They also reported that stream penetration increases the influence zone.

![Figure 3. Flux responses during pumping.](image)

### 2.1.1 Modelling at complexity levels 1–2; analytical modelling

There are numerous analytical solutions for stream depletion derived for a variety of conceptual systems for pumping. Theis (1941) obtained the first unsteady solution for the stream depletion due to abstractions from a fully penetrating well. Glover and Balmer (1954) re-wrote the Theis solution in terms of the complementary error function. This basic stream depletion model assumes a single-layer homogeneous aquifer where the stream fully penetrates the underlying aquifer (i.e., stream depth equal to aquifer thickness). Singh (2000) showed that the two analytical solutions are the same in the mathematical sense by solving the integral in the Theis solution to arrive at the solution of Glover and Balmer. Cooper and Rorabaugh (1963) provided solutions for the unit step response function for a finite aquifer. Hantush (1965) considered a streambed lined with a less permeable aquitard and Hantush (1967) developed an analytical solution for depletion of flow in right-angle stream bends. Jenkins (1968) and Wallace et al. (1990) showed how superposition and time translation are used with the Theis solution to obtain solutions for more general pumping schedules. Hall and Moench (1972) showed that the convolution relation can be used to simplify the mathematics of stream aquifer interactions, permit greater generality by allowing for flood pulses of arbitrary shape, and evaluate quantitatively flow in and out of the aquifer. Zlotnik and Huang (1999) modelled stream depletion from a partially penetrating, finite width stream with a semi-permeable bottom. Hunt (1999) proposed an analytical solution that incorporates the effects of finite stream width, shallow stream penetration, and low-permeability streambed based on the Dupuit assumptions. From these studies it can be concluded that:

- An increase in stream width results in increasing stream depletion.
- The effect of increase in the stream width becomes limited due to increase in the length of the groundwater path under the stream.
- When the stream leakance is large, the stream depletion curves approach the case of a fully penetrating stream.

Hunt (1999) formulated an analytical solution for stream depletion and concluded that the solution by Hantush (1965) can be used to describe flow depletion from a slightly penetrating streambed in an infinite aquifer by modifying the stream leakance. Singh (2000) derived expressions for volume of stream depletion for pumping and recovery. Moench and Barlow (2000) presented Laplace transform step-response functions for several cases of transient,
hydraulic interaction between a fully penetrating stream and a confined, leaky, or water table aquifer. The various aquifers may be semi-infinite or finite in width and may or may not be connected with the stream through a semi-impervious stream bank. Chen (2001) presented a particle tracking technique to evaluate the transport of the leaked stream water in a nearby aquifer. Darama (2001) developed an analytical solution for stream depletion during non-uniform cyclic pumping using the principle of superposition. It was concluded that as the value of pumping distance between the well and stream decreases, the error produced by neglecting the effect of the clogging layer of the streambed may be as high as 90%. Zhan and Butler (2005) presented mathematical derivations of drawdown and stream depletion produced by pumping in the vicinity of a finite-width stream of shallow penetration. Fox et al. (2002) modified Hunt’s (1990) analytical solution through the use of Heaviside functions to allow a distributed recharge flux through the entire stream width rather than modelling the stream as a line source. Butler (2003) presented mathematical derivations of expressions for the total pumping-induced leakage entering an aquifer. Hunt (2003) obtained a solution for flow to a well in a semi-confined aquifer when the well is beside a partially penetrating stream with zero width (distance between and stream and well is large relative to stream width). Bakker and Anderson (2003) presented an explicit analytical solution for steady, two-dimensional groundwater flow to a well near a leaky streambed that penetrates the aquifer partially. The solution was used to investigate the interaction between groundwater and surfacewater in the stream, the effects of pumping on the opposite side of the stream, and the effects of the leaky streambed on the capture zone envelope of the well. Hunt (2003) derived unsteady solutions for stream depletion allow water resource managers to devise well pumping schedules that can reduced the harmful effects of stream depletion to acceptable levels. Singh (2003) presented equations for the rate and volume of stream flow depletion applicable for intermittent pumping cycles; graphs suitable for engineering applications for siting of wells have also been presented. Chen (2003) presented analytical solutions that can be used to evaluate stream infiltration and base flow reduction by groundwater pumping in nearby aquifers. Critical time, infiltration reach, and travel times were calculated to determine the hydraulic connectivity between the well and the stream. Zlotnik (2004) stated that in leaky aquifers, stream depletion may only partially support groundwater withdrawal from a pumping well. The maximum stream depletion rate (MSDR) was defined as a fraction of the pumping rate supplied by stream depletion. It is shown that a large contrast of hydraulic conductivity between the pumped aquifer and underlying bed ($10^4 - 10^6$) may not be a sufficient criterion for considering the aquifer base as an aquiclude. The proposed methods can be used for approximate assessment of stream depletion rates and the MSDR in alluvial valleys with different sources of recharge and discharge. Rassam et al. (2005a) presented an analytical solution for depletion of flow for a U-shaped meandering river. Knight et al. (2005) arrived at the solution of Glover and Balmer (1954) but used it to assess the impacts irrigation on stream salinity; i.e., used the solution in an inverse manner to model recharge/discharge. They also provided analytical solutions for bounded aquifers. Zhan and Butler (2005) presented mathematical derivations of semi-analytical solutions for pumping-induced drawdown and stream depletion in a leaky aquifer system. Hantush (2005) derived closed form solutions for channel flow and stream-aquifer interactions; it relates channel reach discharge, stream aquifer exchange rates, and associated flow volumes to hydrologic processes and regulatory and management control measures. Di Matteo and Dragoni (2005) obtained an empirical equation for stream depletion for the case of a stream that partially penetrates the aquifer and a pumping well that is screened over a portion of the aquifer. The equation is used to minimise the quantity of water coming from a partially penetrating river, which may be polluted. Singh (2005) and Singh (2006) provided an approximate analytical solution for obtaining pumping induced rate and volume of stream flow depletion, which can account for any variation in pumping discharge.
2.1.2 Modelling at complexity level 3; numerical modelling

Sophocleous et al. (1995) used MODFLOW (McDonald and Harbaugh 1988) to assess the predictive accuracy of stream-aquifer solutions; the three most important factors tested, which relate to the multidimensional nature of the aquifer flow conditions, are: (1) streambed clogging as quantified by streambed-aquifer hydraulic conductivity contrast; (2) degree of stream partial penetration; and (3) aquifer heterogeneity. They concluded that serious effort should be made to quantify with reasonable accuracy the streambed-aquifer hydraulic conductivity contrast. Stream partial penetration significantly reduces stream depletion. The effects of large-scale heterogeneity were investigated; for a horizontally layered, the analytical solution overestimated stream leakage whereas it underestimated it for the case of transverse heterogeneity. The relative importance of the various assumptions involved in the derivation of the analytical solution results in the following ranking order of significance:

- Streambed clogging.
- Degree of stream partial penetration.
- Aquifer heterogeneity.
- Nonequilibrium losing stream.
- Aquifer storativity.

Butler et al. (2001) compared analytical solution and numerical predictions to conclude that the conventional assumption of fully penetrating stream will lead to a significant overestimation of stream depletion of more than 100% in many practical applications; the degree of overestimation will depend on the value of the stream leakance parameter and the distance from the pumping well to the stream. They also concluded that aquifer width must be on the order of hundreds of stream width before the assumption of a laterally infinite aquifer is appropriate for stream depletion calculations. Conrad and Beljin (1996) specified the criterion that streambed effects cannot be neglected if the streambed hydraulic conductivity is more than one order of magnitude lower than the aquifer hydraulic conductivity.

Kollet and Zlotnik (2003) concluded that aquifer heterogeneity caused inconsistencies in parameter estimates and stream depletion predictions. They stated that the concept of streambed conductance appears to be a too simplified approximation of the hydraulic connection between the stream and the aquifer. Incorporating large-scale aquifer heterogeneity via piece-wise homogeneous models that are based on sedimentologic concepts of the stream aquifer system under consideration might decrease the uncertainty in stream depletion rate predictions.

Chen and Shu (2002) used numerical modelling techniques to simulate stream aquifer interaction with seasonal groundwater pumping. They concluded the following:

- For a shallow penetrating stream with a low streambed conductance, base flow reduction accounts for a significant percentage of the total stream depletion. Its residual effects in post-pumping can last a long time and may continue into the next pumping season for areas where recharge is nominal.
- In contrast, the contribution of the induced stream infiltration to the total stream depletion is much smaller, and its effects often become negligible shortly after pumping was stopped.
- Residual effects would be significantly lower by taking into account areal recharge.
- Base flow reduction can account for more than 90% of the total depletion for some stream aquifer conditions where a stream is characterized by low conductance.

Rotting et al. (2006) presented a methodology for the design and execution of stream stage response tests and their joint numerical interpretation with pumping tests. The advantage with the proposed technique is that it can account for aquifer heterogeneity and resolves aquifer diffusivity into transmissivity and storage coefficient.
2.2 Surfacewater driven processes

2.2.1 Overland flow and through-flow

The stream flow response to precipitation is dependent on the flow pathways of the watershed, which include direct (on-stream) rainfall, overland flow, through-flow or shallow subsurface flow, and groundwater flow. Overland flow is described as the water that flows over the ground either as quasi-laminar sheet flow or as flow anastomosing in trickles and rivulets, while through-flow (or interflow) refers to subsurface flows that travel laterally to streams through unsaturated soil and in ‘perched’ saturated zones (Ward and Robinson 2000). The accumulation of channel precipitation, overland flow and rapid throughflow is termed ‘quickflow’.

Overland flow is influenced by spatial variation in topography, vegetation, soil characteristics and geology (Jenkins et al. 1994). Overland flow occurs when rainfall intensity exceeds the infiltration capacity of the soil (e.g. Horton 1933) or when soil becomes saturated (e.g. Hewlett 1961), and depression storages are filled, leading to downhill flow. Through-flow is essentially that part of the subsurface discharge to a stream that is not from the regional groundwater system. Through-flow is intensified by low permeability horizons in the soil profile or by high vertical-to-horizontal anisotropy in the soil permeability (Ward and Robinson 2000), and is particularly important in temperate forested environments (Hutchinson and Moore 2000) or steep hill slope catchments (e.g. Moore and Thompson 1996).

In some cases, it may be somewhat challenging to distinguish between groundwater discharge, through-flow, overland flow and/or quickflow, as indicated by the study of Uchida et al. (2006) that explored processes of subsurface drainage occurring in two well-instrumented and similar hill slope catchments. Bari and Smettem (2004, 2006) were able to reproduce stream flow observations using a calibrated daily water balance model that distinguished between various subsurface storages and simulated lateral flows in both groundwater and soil stores. Bormann et al. (1999) used physically based models to separated interflow, surface runoff and groundwater flow to show that under certain constraints, small scale models can be aggregated to simulate large scale behaviour of catchment hydrological processes. Wardlaw et al. (1994) were also able to separate interflow, groundwater flow and overland flow in their calibrated integrated-catchment management model of the Allen River. Kunkel and Wendland (2002), on the other hand, combined interflow and surface runoff in their study of the long-term water balance of the river Elbe. In studies that characterise hydrological processes using hydrograph separation techniques, combining all subsurface processes into a single ‘base flow’ component is commonplace (e.g. Muller et al. 2003). Where quantification of interflow is necessary, models such as TOPMODEL (Moore and Thompson 1996; Walter et al. 2002) can be used to predict shallow interflow of perched groundwater.

In deterministic hydrological models, overland flow hydrographs are usually generated using lumped or semi-distributed (i.e. lumped at the sub-basin scale) catchment models that are calibrated to stream gauge data, where available (Jenkins et al. 1994). An alternative to deterministic modelling is the stochastic approach, which comprises time series analyses of rainfall and stream flow records to investigate hydrological event frequency (e.g. Shaw 1988). Various methods of catchment runoff estimation exist, and these include:

- Numerous variations of the unit hydrograph approach (e.g. Clark 1945; Shaw 1988; Cordery 1993; Maidment et al. 1996)
- Catchment storage routing methods - e.g. the Nash Model (Nash 1957), the RORB model (Laurenson et al. 2006), XP-RAFTS (XP Software 2000), AWBM (Boughton and Chiew 2003), the Sacramento model (Burnash et al. 1973)
- Distributed models – e.g. MIKE SHE (Refsgaard and Storm 1995) and CATFLOW (Zehe et al. 2001)
- Various others (e.g. Lischeid and Uhlenbrook 2003).
2.2.2 River flow attenuation

Storage effects in rivers influence the passage of flow events, as evidenced by the re-
distribution of stream flow hydrographs at successive points along a river. Hydrograph
attenuation is characterised by a reduction in peak flows and an increase in time lag. It
depends on the volume of the stream flow event compared to the volume of storage, and also
on the physical attributes of the system such as length, slope and hydraulic resistance (Pilgrim
1987). The determination of the timing and magnitude of flow at points along a watercourse
from known upstream hydrographs is referred to as ‘flow routing’. In hydrologic studies, flow
routing generally assumes a lumped representation of the system, whereby flow hydrographs
are calculated only for particular locations (Dooge 1986; Chow et al. 1988). Hydrological
routing methods solve the continuity equation using simplifying relationships between the
channel storage ($S$) and flow ($Q$) or the time derivative of flow ($dQ/dt$). Methods include the
Level Pool method, the Muskingum method, the Linear Reservoir model and the Modified
Pulse method (Chow et al. 1988, Pilgrim 1987).

Studies of riverine hydraulics, being distinct from riverine hydrology, alternatively adopt
distributed-system approaches, and solve various simplified forms of the Saint-Venant
equations. Hydraulic routing methods differ by the degree of simplification of the momentum
equation, and approaches include the kinematic wave, diffusion wave and dynamic wave
models (Chow et al. 1988).

In hydrological or hydraulic modelling studies that incorporate groundwater-surface interaction,
the simulated attenuation of stream flow events can influence predicted groundwater–
surfacewater exchange rates, which are usually linked to hydraulic gradients across the
stream-aquifer interface. That is, stream-aquifer hydraulic gradients depend on stream water
levels, which are a function of stream flow, and therefore flow event attenuation. For this
reason, in recent studies of stream-aquifer interaction, models such as MODHMS
(HydroGeologic-Inc. 2001) that incorporate stream flow attenuation are preferred over models
like the STR package (Prudic 1989), in which attenuation is neglected (e.g. Werner et al.
2006).

2.2.3 In-stream storages and reservoir operation

In terms of stream flow events, increasing the in-stream storage enhances the river flow
attenuation, i.e. reduces peak flows and increases time-lags. The simulation of the effects of
in-stream storages on hydrographs can be estimated using the above-mentioned Level Pool
method or various other methods as described by Pilgrim (1987). Determining the impact of
in-stream storages on the low flow hydrology (including stream-aquifer interaction), rather than
the flood hydrology, of a river is less straightforward, and is highly dependent on the regulation
of dam/weir releases, and the characteristics of the impoundment (Smakhtin 2001). The
impact of in-stream development on the interaction between groundwater and surfacewater in
the Los Alamos Canyon was explored recently by Levitt et al. (2005). In some cases, in-
stream storages are constructed to enhance recharge to the adjacent aquifers; the resulting
fluxes are termed ‘artificial recharge’ (Abu-Taleb 2003; Bouwer 2000).

A general discussion of reservoir operation is given by Shaw (1988 p. 464–476) and Eichert et
al. (1982, p. 99–105). Extensive research into the optimal operation of reservoirs exists, and
the reader is directed to the review by Labadie (2004) for an overview. Storage releases may
potentially control exchange flows between downstream river reaches and adjoining
groundwater systems (e.g. Springer et al. 1999). In many cases, analyses of reservoir
operation necessarily incorporate reservoir-groundwater interactions, and groundwater
recharge resulting from leakage from impoundments may significantly modify the subsurface
hydrology and subsequently effect river-aquifer interactions (e.g. Dawoud and Allam 2004;
2.2.4 Off-stream storages

All dams are subject to leakage, either through the foundation and abutments, or through the embankment itself in the case of earth and rock-fill dams (US Army 1993). Few studies have explored the impact of leakage from off-stream storages on both groundwater hydrology and groundwater–surfacewater interaction, although dynamic modelling of reservoir-groundwater interactions is described by Ruiz and Rodriguez (2002). Surfacewater impoundments may be used to augment groundwater storage through the artificial recharge that results from reservoir leakage. Bouwer (2000) provides a discussion of artificial recharge from both in-stream and off-stream storages.

In some cases, off-stream storage may take the form of wetlands. Wetlands have many hydrologic roles, and the groundwater–surfacewater interactions associated with wetlands are often complex (Ward and Trimble 2003). Wetland hydrology is discussed extensively by Ward and Trimble (2003 p. 314–317), DEQ (2001 p. 9–16) and Carter (1996). Reviews on wetland hydrological research are given by Price et al. (2005), Bedford (1996), Koreny et al. (1999) and Hey et al. (1994).

2.2.5 Bank storage

During inter-storm periods there is a streamward hydraulic gradient in gaining streams that maintains groundwater discharge into them. Stream water levels rise in response to runoff and, in most cases, results in the reversing of the hydraulic gradient, which induces a net flux into the floodplain. This water is temporarily stored in the floodplain and is slowly released back to the stream when the stream water level drops and the gradient towards the stream is re-established. This phenomenon is referred to as bank storage.

Significant bank storage occurs when: (1) a stream reach is subject to stage increases; (2) bank materials have a high hydraulic conductivity; and (3) sufficient volumes of permeable bank material provide storage.

Downstream reaches are more favourable to bank storage than headwater reaches (Kondolf et al. 1987). They are subject to high stage increases because they have greater drainage areas that produce large flood peaks, and are more likely to be flanked by alluvium that provide significant storage relative to stream flow. The variability of the hydraulic conductivity of bank materials is what prevents bank storage from being a ubiquitous phenomenon. High-gradient, straight-to-braided streams, would be more likely to produce alluvial fill of high hydraulic conductivity than would low-gradient, meandering streams (Kondolf et al. 1987). Rivers dominated by infrequent events might also be more likely to produce coarse deposit because of the high competence of flood flow. Watershed lithology is another important factor. Rivers that drain granite terrane are more likely to have coarse-grained alluvium than are rivers that drain shales, because granite rocks weather sands and gravel, the shales to mud.

The morphology of a riverbed is determined by geological conditions, by suspended sediment concentration and particle size distribution, and by the flow velocity of the river. Sedimentation of suspended solids reduces the hydraulic conductivity of the stream aquifer interface. An additional clogging mechanism is caused by mass flux from the river into the aquifer. Particulate organic matter intensified clogging of the riverbed and compromised the sustainability of riverbank filtration (Schubert 2002). Riverbed conductance plays an important role in dampening the propagation of flood waves entering the aquifer. Pinder and Sauer (1971) conducted a modelling study and found out that the importance of bank storage depends most heavily on the hydraulic conductivity of bank material and much less on the width of the alluvial aquifer. This is consistent with other observations that bank recharge is concentrated near the stream bank (Todd 1955).

Floodplains mitigate against extreme hydrologic phenomena by storing substantial volumes of water which otherwise increase flood volumes. During a storm event, bank storage diminishes and delays flood peaks (Cooper and Rorabaugh 1963, Moench et al. 1974; Hunt 1990).
Drainage of water into the channel sustains stream flow between storm events (Todd 1955). Bank storage contribution to base flow may be important on alluvial rivers with highly permeable bank materials (Kondolf et al. 1987). The conditions likely to sustain base flow involve a balancing between yield and duration of flow. There must be sufficient discharge from bank storage to maintain a free-flowing stream during drier seasons, but drainage must not be so rapid so to disappear before precipitation replenishes the stream and bank storage. Drainage from the floodplain with a drop in channel water level occurs over a period of days in gravel, weeks to months in sand, years in silt, and decades in clay. Channels flowing through gravel floodplain are likely to become losing reaches; thus supplies of upstream waters are critical to maintain in-stream flows. Sandy narrow streams behave similarly. Sandy wide floodplains provide a good yield of water and duration of flow for sustaining in-stream flow but not through prolonged drought periods (Whiting and Pomeranets 1997). On regulated rivers, if base flow from the upper watershed is impounded, bank storage may be an important source of base flow to downstream alluvial reaches (Kondolf et al. 1987). Bank storage has a significant influence on base flow dynamics in response to a flood wave travelling in a channel with highly permeable riparian aquifers; a 5-day flood wave with the maximum rise of 2 m can generate base flow of up to 46 m$^3$/d per metre length of stream channel (Chen et al. 2006). On the Carmel River, California, bank storage contribution was detected two months after the last peak flow of the rainy season, during a period of critical importance to trout. However, in extremely wet years, the sustained base flow from the upper basin overwhelms the more transient bank storage contribution (Kondolf et al. 1987).

Irrigation on the floodplain and withdrawal of groundwater from the alluvial aquifer significantly affects bank storage; e.g., pumping of the alluvial aquifer can lower the alluvial water table adjacent to the stream so that stream flow is locally influent to the groundwater (Kondolf et al. 1987). This can provide additional storage for recharge from flood flows.

The bank storage phenomenon also occurs within ephemeral stream systems especially those with floodplains having the potential to perch water. Based on analysis of 7-months of field data, (Rassam et al. 2006b) have showed that the volumes of water stored in the banks of a perennial stream and one of its ephemeral tributaries were similar but responded differently to rainfall patterns due to different storage mechanisms in the two stream systems.

Bank storage affords opportunities for transforming and immobilizing pollutants and nutrients (Peterjohn and Correll 1984). Discharge from bank storage can control water quality during base flow periods (Squillace 1996). Lamontagne et al. (2005) reported that the water and nitrogen budgets of an alluvial aquifer were mostly driven by cycles of bank recharge and discharge between the stream and the alluvial aquifer.

Bank filtration employs a natural filtration process of surfacewater on its flow path from the river to a production well (Wett et al. 2002). In the lower Rhine region, riverbank filtration for water supply has been employed for 130 years (Schubert 2002).

**Level 1 modelling; conceptual lumped approach**

Cooper and Rorabaugh (1963) developed an empirical expression for the groundwater outflow for a finite width, uniform, homogeneous and isotropic aquifer drained by an intersecting stream:

$$Q = Q_0 \exp(-at)$$

(1)

Where $Q$ is discharge, $t$ is time, and $a = \pi^2 T/4 S L^2$ is a recession parameter (with $T$ and $S$ representing aquifer transmissivity and storage coefficient, respectively).

Rassam et al. (2005b) formulated a model for riparian denitrification during bank storage using a similar simplistic approach to model the time delays associated with bank storage. Rassam et al. (2006a) used the same approach to estimate $^{15}$N and $^{18}$O fractionation resulting from denitrification during bank storage.
Level 2 modelling; one-dimensional approach

The interaction between a stream and the groundwater in an adjacent unconfined aquifer has been considered by many authors, including Glover and Balmer (1954), Todd (1955), Venetis (1970), Hall and Moench (1972), Hantush (2005), and Knight et al. (2005). Most studies have used the linearized Boussinesq equation to describe changes in the groundwater level. In this case standard solutions of the heat equation from Carslaw and Jaeger (1959) can be used to derive impulse response functions for the effects of groundwater recharge or pumping on lateral flux into or out of the stream, or the effect of changing river levels on groundwater heights. Hall and Moench (1972) studied an aquifer with a thin semipervious layer between stream and aquifer, and authors such as Hantush (1967), Venetis (1970), Zlotnik et al. (1999) and Moench and Barlow (2000) have modelled some kind of non-uniformity in the spatial domain.

For example, the head response $h$ in an aquifer due to a unit rise in stream level is

$$ h(x,t) = \operatorname{erfc} \left[ \frac{x}{2 \sqrt{D t}} \right], \quad h(0,t) = 1, \quad t > 0, $$

(2)

where $D$ is the aquifer diffusivity, $t$ is time, and $x$ is the distance between a point in the aquifer and the stream bank (Carslaw and Jaeger 1959 p. 63). It is worthwhile noting that Equation (2) is identical to the well-known Glover and Balmer (1954) solution used to evaluate stream depletion.

When the input function is more complicated than a single pulse or step function of time, combinations of pulse or step functions can be used for the input, as in (Knight et al. 2005). For inputs which are general functions of time the output is given as a convolution of the appropriate impulse response with the input, as used by Venetis (1970), Barlow et al. (2000), Hall and Moench (1972), Singh (2004a), Singh (2004b) and others. In general the convolution integral must be evaluated numerically in the time domain, and care must be taken to avoid loss of accuracy in certain parameter ranges. If the input function is given as a set of measured values at certain values of time then some form of interpolation must be used to calculate input values at other times. Another approach is to use a combination of many finite pulse inputs, for each of which an analytical solution is known. (Singh 2004a) investigated the use of ramp function inputs, and found that they were superior to pulse inputs.

![Figure 4. Head response due to basis spline input.](image)

Knight and Rassam (2007a) proposed the use of cubic basis splines to show how to calculate explicitly the analytical response of groundwater levels in an unbounded aquifer to random changes in stream water level. The proposed technique overcomes the difficulties and uncertainties arising from the numerical evaluation of the convolution integrals associated with...
other methods. The head response at two points in the aquifer due to a cubic basis spline stream pulse is shown in Figure 4; the response varies linearly with aquifer diffusivity ($D$) and non-linearly with the distance from the stream ($x$).

Knight and Rassam (2007b) derived an analytical solution for the flux response to a cubic basis spline stream pulse, which is shown in Figure 5. It is shown that the time scales for bank storage are strictly related to the shape of the input stream-stage fluctuations with aquifer properties affecting only the magnitude of the fluxes and volumes of bank stored water. The technique is used to evaluate stream bank fluxes and bank storage during a random flood wave.

![Figure 5. Characteristic flux responses to a cubic basis spline input; $C$ is aquifer conductance and $\phi$ is storage coefficient.]

**Level 3 modelling; multi-dimensional approach**

Whiting and Pomeranets (1997) developed a 2D model for bank storage (WaTab2D) with non-symmetrical channel banks, non-uniform hydraulic geometry, and non-zero boundary fluxes.

Chen and Chen (2003) conducted a numerical modelling study using MODFLOW and MODPATH to trace the pathlines on the infiltrated stream water and determine the size of the bank storage zone. They suggested that the storage zone (defined as the part of aquifer where groundwater is replaced by stream water during flood) reflects the hydraulic connectivity between stream and aquifer. Chen and Chen (2003) investigated the effects of stream aquifer fluctuations, aquifer properties, the hydraulic conductivity of streambed sediments, regional hydraulic gradient, and recharge and evapotranspiration rates on stream aquifer interaction. They concluded that bank storage solely caused by stage fluctuations differs slightly between losing and gaining streams.

The effects of bank storage have been incorporated into various modelling experiments. Griffiths and Clausen (1997) modelled the falling limb of the outflow stream hydrograph from a natural basin. The recession formula accounts for depression (water stored in lakes, ponds, marshes, and other depressions on the ground surface), detention (storage in overland flow and is analogous to channel storage in a stream), snow and ice, channel, channel bank (bank storage), aquifer and cavern (karst terrain) storage, and evapotranspiration. Birkhead and James (2002) modified the Muskingum procedure to explicitly account for the interaction between channel flow and bank storage in rivers with permeable banks of varying hydraulic conductivity. They implemented the method for a 4.6 km reach of the Sabie River in South Africa and found out that bank storage accounts for up to 40% of the temporal reach storage, indicating the significance of bank storage within this system.
2.2.6 Over-bank flooding

Two important components contributing to the water budget of an alluvial valley aquifer are recharge and stream-aquifer interaction (infiltration and exfiltration of water across the streambed). Overbank flooding is a key hydrologic process affecting riparian water table dynamics and ecological processes such as biogeochemical cycling and plant diversity (Naiman and Decamps 1997). Overbank flooding typically occurs for a few days to weeks on average after years for most natural rivers (Wolman and Leopold 1957). Riparian soil water and groundwater recharge can be greater during overbank flooding than from river aquifer interactions or precipitation. For example, during a five-year study in south central Ohio, Workman and Serrano (1999) found that significant overbank flow during and shortly after large flood events overshadowed all instances of recharge that could be attributed to infiltrating water. Direct recharge attributed to overbank flow during two large flood events accounted for 65% of the total recharge computed.

River flows determine the distribution of patterns of channels, back-swamps, marshes, and tributaries that make up the floodplain. These floodplain wetlands also include freshwater and saline lakes, anabranches, billabongs, lagoons, overflows, swamps, and waterholes in Australia (Kingsford 2000). Arrival of water in a floodplain wetland sets off a dynamic ecological processes and interactions among a wide range of species. Kingsford (2000) reported that Australian floodplain wetlands are sites of extraordinary biological diversity with abundant and diverse populations of waterbirds, native fish, invertebrate species, aquatic plants, and microbes.

Dams on rivers store water hence reducing the amount of water that could potentially reach floodplain wetlands thus affecting their ecology. Loss of connectivity to the river changes aquatic systems to terrestrial ecosystems. Substitution of a variable-floodplain pattern with a permanent one, and loss of wet-dry cycles, has lasting ecological effects. Reduction of flood peaks resulting from dam constructions on rivers, alters the floodplain hydrology by having less groundwater replenishment and diminishing the hydraulic connectivity between the river and the floodplain (Girard et al. 2003). The reduction in the requisite discharge for overbank flow, together with a reduction in the peak discharge regime, shows that recent flooding along the mainstem Skokomish River, Washington, is due to aggradation within the channel that reduced channel conveyance (Stover and Montgomery 2001). In contrast, Westbrook et al. (2006) found that construction of beaver dams on the Colorado River greatly enhanced the depth, extent, and duration of inundation associated with floods and also elevated the water table during both high and low flows. This created and maintained hydrologic regimes suitable for the formation and persistence of wetlands.

Whigham and Young (2001) proposed a simple water movement model to allow prediction of frequency and duration of inundation of different floodplain environments. Water bodies are defined as a one-dimensional storage representing the quantity of water storage per unit time. They are filled and drained by conceptual pipes that have a limited capacity and a threshold for water release. The framework is used in the development of an environmental flows decision support system.

2.3 Groundwater–surfacewater driven processes

2.3.1 Groundwater–surfacewater interactions in wetlands

Wetlands include swamps, marshes, billabongs, lakes, saltmarshes, mudflats, mangroves, coral reefs, fens, peatlands, or bodies of water — whether natural or artificial, permanent or temporary. Water within these areas can be static or flowing, fresh, brackish or saline (Department of the Environment, Water, Heritage and the Art; http://www.environment.gov.au/water/environmental/wetlands/about.html). Cowardin et al. (1979) defined wetlands as ‘land transitional between terrestrial and aquatic systems where the water table is usually at or near the surface or the land is covered by shallow water’. They added that wetlands must have one or more of the following attributes: (1) at least periodically,
the land supports predominantly hydrophytes; (2) the substrate is predominantly undrained hydric soil; and (3) the substrate is non-soil and is saturated with water or covered by shallow water at some time during the growing season of each year. Wetlands have complex hydrological interactions because they are subject to periodic water level changes (Winter et al. 1998). Hydrologically, wetlands are characterized by the periodic excess of water inflow over outflow that provides a saturated substrate. Wetlands are present in climates and landscapes that cause groundwater to discharge to land surface or that prevent rapid drainage of water from the land surface. Wetlands do not always occupy low points and depressions in the landscape. Wetlands can be present on slopes (such as fens where groundwater discharge at break of slope) or on drainage divides (such as bogs). Hydroperiod is a term commonly used in wetland science that refers to the amplitude and frequency of water level fluctuations. It affects all wetland characteristics, including the type of vegetation, nutrient cycling, and the type of invertebrates, fish, and bird species present (Winter et al. 1998).

Drexler et al. (2004) described wetland types as: non-tidal freshwater marshes, tidal freshwater marshes, freshwater swamps, mangroves, peat-lands, riparian wetlands, and salt marshes. DeBusk (1999) grouped wetlands according to their position in the landscape as follows:

- Depressional wetlands, as the name implies, form in a depression in the landscape and are not directly associated with rivers and lakes. Hence, they do not play a critical role in the hydrology of the watershed. Some are effectively isolated from direct interaction with groundwater, and are sometimes referred to as ‘perched wetlands’. The water budget of the latter is mostly exclusively controlled by rainfall and evapotranspiration.
- Riparian wetlands (also referred to as floodplain wetlands) are found in low-lying regions adjacent rivers and streams that are periodically subjected to overbank flooding. Their hydraulic connection to rivers and streams implies that they are of major importance in regional hydrology. They intercept surface as well as sub-surface flow and thus function as buffers for the river system.
- Fringe wetlands adjacent to lakes and estuaries are also hydrologically connected with uplands and aquatic ecosystems and they have less hydrologic influence than riparian wetlands.

Since wetlands cover approximately 6% of the world’s land area, their effect on the water cycle is significant (Bullock and Acreman 2003). Winter et al. (1998) stated that the main drivers of wetlands hydrology are: precipitation, evapotranspiration, and interactions with surfacewater and groundwater. The main inputs of water into wetlands include: precipitation, influent river seepage, overbank floods, and groundwater inflow; outflows include: evapotranspiration, effluent river seepage, surface runoff, and groundwater outflow (Andersen 2004). The hydrology of wetlands is largely controlled by their position in the groundwater flow system (Todd et al. 2006); the interactions of wetlands with groundwater and surfacewater are also affected by the geologic characteristics of their beds, and their climatic settings (Winter 1999).

Many wetlands are associated with permeable deposits where the influx of regional groundwater plays a critical role in maintaining relatively constant surface saturation (Waddington et al. 1993). Other wetlands are perched above regional groundwater flow systems in topographic depressions in geologic materials with low permeability. Wetlands that occupy depressions in the landscape have interactions with groundwater similar to lakes and streams.

Climate effects on wetland function include temperature and precipitation. Temperature regimes control the rates of important biological processes, such as those involving organic matter decomposition and consequently, accumulation of peat in the wetland. Precipitation has a substantial effect on wetland hydrology; also significant is the timing and pattern of rainfall, which affect the frequency and duration of flooding (the hydroperiod), as well as vegetation types (DeBusk 1999). There is strong evidence that wetlands evaporate more water than other land types and reduce the flow of water in downstream rivers during dry periods (Bullock and Acreman 2003). Crundwell (1986) concluded that wetland evapotranspiration is generally greater than open water evaporation (at least during spring
Review of groundwater–surfacewater interaction modelling approaches

Many wetlands exist because they overlie impermeable soils or rocks where there is little interaction with groundwater. However, Raisin et al. (1999) highlighted the importance of groundwater in the water balance and nutrient budget for an Australian wetland; their investigation confirmed that a wetland in the Kiewa Valley in north-eastern Victoria was a significant discharge area where groundwater accounted for 97% of the surfacewater and 50% of the total nitrogen leaving the system. A study of floodplain wetlands in Nigeria has shown that recharge through the annually inundated floodplains is the source of a groundwater mound (Goes 1999). The presence or absence of a uniform clay cover in the floodplain determines whether the annual floods recharge the groundwater and if the shallow aquifer is confined or not. There are conflicting reports in the literature as to whether wetlands reduce or increase recharge to groundwater (Bullock and Acreman 2003), which highlights the site specific nature of groundwater–surfacewater interactions.

Whether hydrological functions of wetlands are considered to be beneficial or not depends upon one’s point of view. For example, ecologists may see evaporation from wetlands as an essential process supporting plant growth, whilst water resource managers may see it as a loss of a vital downstream resource. Most floodplain wetlands reduce or delay floods, with examples from all regions of the world, however, some headwater wetlands might increase flood peaks (Bullock and Acreman 2003). The role of riparian wetlands in improving stream water quality has been highlighted by many researchers, e.g. Burt et al. (2002). They are generally considered to have the most important water quality role in watersheds, due to their strategic location between upland and aquatic ecosystems, they have a high level of biodiversity, and provide critical habitats for many plants and animals. Of particular significance to downstream water quality are riparian wetlands associated with low-order streams because of the significance of their cumulative impact on water quality (Rassam et al. 2005b; Rassam et al. 2006b; and Angier et al. 2005).

Level 1; conceptual modelling

Vidon and Hill (2004) presented a conceptual model of stream riparian hydrology based on the examination of eight riparian sites in Canada. The model emphasizes the importance of upland permeable sediment depth as an indicator of aquifer size and topography of the riparian zone and adjacent upland for the identification of different hydrologic categories of riparian zones.

Rassam et al. (2005b) presented conceptual models for surfacewater groundwater interaction in riparian zones belonging to ephemeral and perennial streams. Those models were used to evaluate the denitrification potential of riparian zones, which is an important process that leads to permanent removal of nitrate from groundwater.

Level 2; water balance models

A water balance is a basic step for making wise and sustainable decisions about wetland hydrology (Peacock and Hess 2004). Studying the water budget of lake-edge or streamside wetlands is difficult because of the difficulty of accurately measuring sheet flow across the wetland surface into or out of the stream or lake (Kadlec 1990). Shallow, horizontal groundwater flow through peat with highly variable hydraulic conductivity, makes the task of understanding the hydrologic budget a difficult one (Ivanov 1981).

Water balance studies, e.g. (Gilvear et al. 1993), have shown that evapotranspiration is the largest hydrological flux through many wetland types during the summer months.

Studying the hydrologic budget of a streamside urban wetland in south central Wisconsin, USA, has shown that precipitation and surface flow comprised over 95% of the inputs to the wetland, while evapotranspiration was the major output (Owen 1995). Owen (1995) also studied the water budget of an urban wetland and concluded that the wetland provided only 21% of the average recharge per unit area (i.e., reduced recharge under the wetland area by 79%). This supports the idea that few wetlands are responsible for significant groundwater...
It was also concluded that the wetland has a large capacity for flood storage; however, there are enormous uncertainties in the water budget of the wetland.

**Level 3; numerical models**

**Unsaturated flow**

Some wetlands are permanently saturated, however, many exhibit partial saturation particularly during the growing season. Bendjoudi et al. (2002) monitored and modelled the hydrologic functioning of a wetland on the Seine River in France identifying and quantifying exchanges between the wetland (especially in the unsaturated upper layer) and its environment. They found out that the river recharges the wetland during the period when evapotranspiration is high. This is the reverse of a natural situation where the alluvial aquifer provides base flow in the rivers during summer, and is partly due to the high flow rates maintained in the rivers by releases from the dams.

The unsaturated zone is potentially very important for wetland ecology for a number of reasons: (1) water content in the root zone affects vegetation species (Bridgham and Richardson 1993) and may be more important to plant growth than water table position (Wheeler 1999); (2) saturation affects aeration, which influences the decomposition of organic matter and nutrient release (Freeman et al. 1997); (3) affects the water budget by affecting hydraulic conductivity; (4) affects storage capacity and percolation (Andersen 2004). A capillary fringe extending to or near the surface maintains near-saturated conditions, which allows very high evapotranspiration. Peat can even under confined conditions, and because of its elastic properties, release considerable amounts of water, which may be replenished by groundwater inflow, influent seepage during high river water level, and percolation during unsaturated conditions (Goes 1999).

**Saturated flow**

Bradley (1996) conducted a numerical modelling study using MODFLOW to study transient water table variations in a floodplain wetland in the UK. The mode had to account for variations of hydraulic conductivities of herbaceous peat, wood peat, and silty clay amounting to four orders of magnitude. Crowe et al. (2004) developed a numerical model for simulating groundwater-wetland interactions and contaminant transport; the model was applied to assess groundwater-wetland interactions and the transport of septic-system-derived contaminants at Point Pelee, Ontario, Canada. Holland et al. (2004) concluded that water level has a direct impact on the residence time distribution, the latter is a critical parameter for modelling and designing treatment wetlands for optimal constituent removal.

**2.3.2 Evapotranspiration**

Potential evapotranspiration (PET) can be generally defined as the amount of water that could evaporate and transpire from a vegetated landscape without restriction other than the atmospheric demand (Penman 1948). Estimating actual evapotranspiration (AET) in moisture stressed settings is usually inferred through the use of a parameterised AET sub-model of the hydrologic model. AET may be modelled in various levels of complexity, which mainly depends on the complexity of the hydrologic system being modelled and the availability of data.

Hydrologic models require spatial and temporal quantification of fluxes of water into, out of, and within the hydrologic system. The significance of AET within the water budget dictates this flux must be estimated in most hydrologic models. It is worthwhile noting that it is the difference between AET and precipitation that provides the ‘available water’ for recharging aquifers and streams, and with any difference of two numbers (particularly if the two are comparable in magnitude), relative error can be amplified, which further supports the need for the best estimates of AET (Sumner and Jacobs 2005).

Quantification of AET is critical to water resource management because it accounts for a large share of the water budget. Worldwide, AET returns about 64% of land-based precipitation to the atmosphere. A case study in the Shashe River in Botswana, it was shown that the amount
of water taken by transpiration is far more than the quantities pumped for water supply (Bauer et al. 2006). In areas of shallow groundwater table, transpiration by near-shore vegetation can intercept groundwater that would otherwise discharge to surfacewater (Winter 1999). Furthermore, it is not uncommon for transpiration from groundwater to create cones of depression that cause surfacewater to seep into the neighbouring aquifer.

Direct measurement of AET provides an opportunity to improve the quality of hydrologic model calibration through reduction in the uncertainty of the AET component of the model in one of two ways: (1) prescribing the AET input in the model to the measured values or (2) comparison of the AET values inferred through the AET sub-model with the measured values, followed by refinement of the conceptualization or parameterisation of the evapotranspiration sub-model.

Here, we will present three alternative methodologies showing how actual evapotranspiration is handled in hydrological models at varying levels of complexity. Usually, the model user specifies potential PET and the model uses some indicator of moisture availability (like water table depth or soil moisture or suction) to calculate AET.

**Level 1; lumped effect of ET on groundwater level**

This approach is relevant for a saturated flow model that accounts for changes in groundwater level due to ET. The traditional approach for modelling AET in a groundwater model such as MODFLOW assumes a piecewise linear relationship between evapotranspiration flux rate and hydraulic head. Bauer et al. (2006) modified SEAWAT (Guo and Langevin 2002), where evaporation and transpiration are treated as two independent diffuse sink terms.

Baird and Maddock (2005) presented a new approach for modelling evapotranspiration based on multiple non-linear, segmented flux curves that reflect the eco-physiology of the plant species in these systems; evapotranspiration flux rate curves set the extinction and saturation extinction depths and define the group’s evapotranspiration flux rates as a function of water table depth relative to the ground surface.

**Level 2; modelling of evaporation and transpiration in a lumped water balance model**

In this approach (and the following one), evapotranspiration is partitioned into evaporation and transpiration and then each of the actual components are calculated separately. The partitioning of evapotranspiration depends on the energy supplied to the crop and soil, and the resistances to transport (Campbell 1985). Partitioning of evapotranspiration may be done simplistically by choosing some arbitrary fraction, or calculated approximately using the leaf area index approach, or more correctly estimated using a complex model such as that presented by Norman and Campbell (1983).

Actual evaporation depends on soil properties and environmental conditions. Even when the soil is covered by vegetation, evaporation is probably at least 10% of evapotranspiration (Campbell 1985). Evaporation is an important component of the water budget especially in dryland farming and desert soils. Actual evaporation and transpiration for this class of models are modelled using empirical relationships.

**Level 3; modelling of evaporation and transpiration in spatially distributed variably saturated flow models**

In this class of distributed models, pressure heads (suctions in the unsaturated zone) are calculated across the entire flow domain. Root distribution functions are used to describe root density. Actual root water uptake can be modelled in various levels of complexity: (1) use algorithms that account for complex processes such as stomatal resistance (e.g., Campbell 1985 chapters 10 and 12), or (2) use empirical relationships that describe actual uptake as a function of the calculated pressure head during a time-marching scheme (e.g. Feddes model in HYDRUS (Šimůnek et al. 1999); RZE1 package in MODFLOW (Thoms et al. 2006)).
Methods for estimating evapotranspiration

There are about 50 methods or models for estimating PET; they give inconsistent values due to their different assumptions and input data requirements, or because they were developed for specific climatic regions (Grismer et al. 2002). Overviews of many of these methods can be found in the literature (e.g. Jensen et al. 1990). These methods can be grouped into several categories, including: empirical, mass transfer, radiation, temperature, combination, and measurement; they have been inter-compared in many studies (e.g. Amatya 1995; Federrer et al. 1996; Lu et al. 2005). It is possible in infer evapotranspiration from pan evaporation data. Grismer et al. (2002) evaluated the relative performance of commonly used non-linear and linear regression equations for prediction of evaporation pan coefficients needed for estimation of reference evapotranspiration from evaporation pan data.

Evapotranspiration rates are estimated by one of the following methods: direct measurement, calculations, hydrological methods, and remote sensing (Pauwels and Samson 2006). The most frequently used direct measurement methods are the Bowen Ratio energy balance and the Eddy Correlation techniques. The most frequently used calculation methods are the Penman-Monteith and the Priestley-Taylor method (Raupach 1991; Shuttleworth 1992; Sumner and Jacobs 2005; Pauwels and Samson 2006). When the resources required for implementing the more sophisticated methods mentioned above are not available, simpler (but not necessarily less costly) alternatives can be used like hydrological methods (Holmes 1984; Bethge-Steffens et al. 2004; Yang et al. 2000; Magaritz et al. 1990). More recently, remote sensing has been used to estimate regional evapotranspiration (e.g. Brunet et al. 1991; Jiang and Islam 1999; Mardikis et al. 2005).
3 Models simulating groundwater–surfacewater interactions

3.1 Groundwater models

Transient groundwater flow through a homogeneous isotropic saturated medium is mathematically modelled using the following partial differential equation, which is formulated by coupling Darcy’s law to the continuity equation that describes the conservation of fluid mass during flow:

\[
\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{S}{K} \frac{\partial h}{\partial t}
\]

(3)

where \( h \) is the hydraulic head, \( S \) is the specific storage, \( K \) is the hydraulic conductivity, and \( t \) is time. When steady-state flow is considered, the right-hand side of Equation 3 becomes zero, which results in the well-known Laplace’s equation. Three dimensional, fully distributed groundwater flow models such as MODFLOW solve Equation 3 numerically to provide the solution \( h(x,y,z,t) \), i.e. hydraulic head at any location in the flow domain at any time. Coupling the unsaturated form of Darcy’s law (where the hydraulic conductivity becomes a function of pressure head) with the continuity equation results in Richard’s equation (Richards 1931), which describes flow in unsaturated media.

The solution of Equation 1 requires sophisticated numerical techniques that are time-consuming. In many cases, however, the flow conditions can be represented or approximated by the two- or the one-dimensional form of Equation 1. For example, the well-known Dupuit-Forchheimer theory reduces the flow system to a horizontal, one-dimensional flow system that is much easier to analyse. Another example for reducing the dimensionality of a flow problem is where the spatial explicitness in the second direction is not required; for example, in a recharge-discharge problem, we can integrate the fluxes along the discharge edge, which results in a one-dimensional term describing flux. However, spatial explicitness may sometimes be required as in the identification of capture zones for pumping wells. In homogeneous, single-layered aquifers, flow can be described with a two-dimensional areal model, but in cases where the geology is complex, a full three-dimensional flow solution is warranted.

In this chapter, we will categorize the complexity of groundwater flow models according to the spatial dimensions that they deal with.

Level 1 models; 1-D flow

The Dupuit-Forchheimer approximation

For flow in unconfined systems bounded by a free surface, an approach pioneered by Dupuit (1863) and advanced by Forchheimer (1930) is often used. It is based on two assumptions: (1) flow lines are assumed to be horizontal and equi-potentials vertical and (2) the hydraulic gradient is assumed to be equal to the slope of the free surface and to be invariant with depth. In effect, the theory neglects the vertical flow component thus reducing a two-dimensional system to one dimension for the purpose of analysis. The technique produces a good approximate solution in cases where the slope of the free surface is small and when the depth of the unconfined flow field is shallow.

Analytical models for stream depletion

Analytical solutions for stream depletion may address one or a combination of the following effects encountered in the field:

- The extent of stream penetration; fully or partially penetrating streams.
- Low hydraulic conductivity streambed.
Stream width effect.
- Bounded aquifers effect (no-flow boundary).
- Partially screened pumping wells
- Handling of stage height variations, i.e., the stream boundary condition.
  - Step-rise.
  - Pumping schedule.
  - Sinusoidal variation.
  - Random variation.

The chronological list of models outlined in Section 4.3 is used to produce Table 1, which groups various models according to the above-mentioned effects.

<table>
<thead>
<tr>
<th>Model developers</th>
<th>Stream penetration</th>
<th>Low-K bed</th>
<th>Stream width</th>
<th>No-flow boundary</th>
<th>Stage height</th>
<th>Other effects</th>
</tr>
</thead>
<tbody>
<tr>
<td>Theis (1941); Glover and Balmer (1954)</td>
<td>Full</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>Step</td>
<td>–</td>
</tr>
<tr>
<td>Cooper and Rorabaugh (1963)</td>
<td>Full</td>
<td>×</td>
<td>×</td>
<td>✓</td>
<td>Step</td>
<td>–</td>
</tr>
<tr>
<td>Hantush (1965)</td>
<td>Full</td>
<td>✓</td>
<td>×</td>
<td>✓</td>
<td>Step</td>
<td>–</td>
</tr>
<tr>
<td>Hantush (1967)</td>
<td>Full</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>Step</td>
<td>Right-angled stream bend</td>
</tr>
<tr>
<td>Jenkins (1968) and Wallace et al. (1990)</td>
<td>Full</td>
<td>×</td>
<td>✓</td>
<td>×</td>
<td>Pumping schedules</td>
<td>–</td>
</tr>
<tr>
<td>Hall and Moench (1972)</td>
<td>Partial</td>
<td>✓</td>
<td>✓</td>
<td>×</td>
<td>Step</td>
<td>–</td>
</tr>
<tr>
<td>Zlotnik and Huang (1999)</td>
<td>Partial</td>
<td>✓</td>
<td>×</td>
<td>×</td>
<td>Step</td>
<td>Confined and leaky aquifer</td>
</tr>
<tr>
<td>Cooper and Rorabaugh (1963)</td>
<td>Partial</td>
<td>×</td>
<td>✓</td>
<td>×</td>
<td>Step</td>
<td>–</td>
</tr>
<tr>
<td>Hantush (1967)</td>
<td>Full</td>
<td>×</td>
<td>✓</td>
<td>×</td>
<td>Step</td>
<td>Right-angled stream bend</td>
</tr>
<tr>
<td>Butler et al. (2001)</td>
<td>Full</td>
<td>×</td>
<td>✓</td>
<td>×</td>
<td>Step</td>
<td>U-shaped meander</td>
</tr>
<tr>
<td>Fox et al. (2002)</td>
<td>Partial</td>
<td>✓</td>
<td>×</td>
<td>×</td>
<td>Step</td>
<td>Partially screened wells</td>
</tr>
<tr>
<td>Rassam et al. (2005a)</td>
<td>Full</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>Step</td>
<td>U-shaped meander</td>
</tr>
<tr>
<td>Knight et al. (2005)</td>
<td>Full</td>
<td>x</td>
<td>x</td>
<td>✓</td>
<td>Step</td>
<td>–</td>
</tr>
<tr>
<td>Di Matteo and Dragoni (2005)</td>
<td>Partial</td>
<td>x</td>
<td>x</td>
<td>✓</td>
<td>Step</td>
<td>Partially screened wells</td>
</tr>
<tr>
<td>Singh (2005); Singh (2006)</td>
<td>Full</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td>Arbitrary; sinusoidal</td>
<td>–</td>
</tr>
<tr>
<td>Knight and Rassam (2007a, 2007b)</td>
<td>Full</td>
<td>✓</td>
<td>x</td>
<td>x</td>
<td>Arbitrary</td>
<td>–</td>
</tr>
</tbody>
</table>

Level 2 models; 2-D flow

**Analytical models:**

CAPZONE (Bair et al. 1992) is an analytical flow model that can be used to construct groundwater flow models of two-dimensional flow systems characterized by isotropic and homogeneous confined, leaky-confined, or unconfined flow conditions. CAPZONE computes drawdowns at the intersections of a regularly spaced rectangular grid produced by up to 100 wells using either the Theis equation or the Hantush-Jacob equation. More details can be obtained from [http://typhoon.mines.edu/software/igwmcsoft/](http://typhoon.mines.edu/software/igwmcsoft/).

Bakker and Anderson (2003) presented an explicit analytical solution for steady, two-dimensional groundwater flow to a well near a leaky streambed that penetrates the aquifer...
The solution was used to investigate the interaction between groundwater and surfacewater in the stream, the effects of pumping on the opposite side of the stream, and the effects of the leaky streambed on the capture zone envelope of the well.

The two-dimensional effects of boundary conditions such as discharge into right-angled stream bends and U-shaped meandering rivers have been addressed by Hantush (1967) and Rassam et al. (2005a).

**Analytic element models**

The analytic element method was developed by Strack (1989). It avoids the discretisation of a groundwater flow domain by grids or element networks. Instead, only the surface-water features in the domain are discretised, broken up in sections, and entered into the model as input data. Each of these stream sections or lake sections are represented by closed form analytic solutions: the analytic elements. The comprehensive solution to a complex, regional groundwater flow problem is obtained by superposition (addition) of a few hundred, analytic elements in the model. Traditionally, superposition of analytic functions was considered to be limited to homogeneous aquifers of constant transmissivity. However, by formulating the groundwater flow problem in terms of appropriately chosen discharge potentials, rather than piezometric heads, the analytic element method becomes applicable to both confined and unconfined flow conditions as well as to heterogeneous aquifers.

GFLOW (Haitjema 1995) is a program based on the analytic element method. It models steady-state flow in a single heterogeneous aquifer using the Dupuit-Forchheimer assumption. While GFLOW supports some local transient and three-dimensional flow modelling, it is particularly suitable for modelling regional horizontal flow. To facilitate detailed local flow modelling, GFLOW supports a MODFLOW-extract option to automatically generate MODFLOW files in a user-defined area with aquifer properties and boundary conditions provided by the GFLOW analytic element model. GFLOW also supports conjunctive surfacewater and groundwater modelling using stream networks with calculated base flow.

While uniquely suitable for groundwater flow modelling at different scales, current generation analytic element models have some limitations. For instance, both transient flow and three-dimensional flow are only partially implemented in analytic element models. Gradually varying aquifer properties cannot be represented in analytic element models. GFLOW does not support multi-aquifer flow. More details can be obtained from [http://www.haitjema.com](http://www.haitjema.com).

SLAEM/MLAEM/MLAEM/2 are a suite of tools for modeling regional groundwater flow in systems of confined, unconfined, and leaky aquifers. SLAEM (Single Layer Analytic Element Model) is the single-layer version of the program, MLAEM/2 (Multi Layer Analytic Element Model) is the two layer version, and MLAEM is the multi layer version. The features currently supported by all programs are the following:

- In-homogeneities in the aquifer parameters, namely, the base elevation, thickness, and hydraulic conductivity. These in-homogeneities are bounded by polygons of straight-line segments and may be nested and have common boundaries.
- Curvilinear features that may be combined to form open and closed strings of the following types:
  - Given head and resistivity (stream bottom).
  - Given head and multi-resistivity (lake and river bottoms that change their behaviour according to the head in the aquifer). This covers the case of portions of lakes and rivers that are dry.
  - Given extraction or infiltration rate per unit length.
  - Constant, but unknown, head and given total discharge (drain with given flow rate).
  - Leaky wall (normal flux is related to the difference in head across the wall and the resistivity of the wall).
  - Impermeable wall.
  - Given flux out of or into one side of the element.
TWODAN has a suite of advanced analytic modelling features that allow the user to model everything from a single well in a uniform flow field on up to complex remediation schemes with numerous wells, barriers, surface waters, and heterogeneities. The aquifer modelled by TWODAN can consist of one or two hydraulically connected layers, it can be confined and/or unconfined, and it can be homogeneous or heterogeneous. The analytic implementation of barrier elements gives much greater accuracy than is possible with numerical methods. TWODAN is also capable of optimizing discharges of steady wells based on specified head and aquifer discharge conditions. In addition, both discharge- and head-specified line sinks are implemented and vertical infiltration or leakage to or from the aquifer can be modelled as uniform or as locally variable. More details can be obtained from http://www.scisoftware.com/products/twodan_overview/twodan_overview.html.

WinFlow is an interactive, analytical model that simulates two-dimensional steady state and transient ground-water flow (both confined and unconfined aquifers) with multiple wells, uniform recharge, circular recharge/discharge areas, and line sources or sinks. The steady-state model is based on analytical functions developed by Strack (1989). The transient model uses equations developed by Theis (1935) for confined aquifers and by Hantush and Jacob (1955) for leaky-confined aquifers, respectively. Any of these functions may be added to a uniform regional gradient for confined and leaky-confined aquifers. More details can be obtained from http://www.scisoftware.com/products/winflow_overview/winflow_overview.html.

Visual BlueBird is a multi-engine graphical user interface for the modelling of single-layer 2-dimensional steady state groundwater flow in heterogeneous aquifers by the analytic element method. The primary numerical engine is the public domain code SPLIT, which supplies such features as particle tracking, capture-zone delineation, and parameter estimation. SPLIT supplies the following high-order elements, which can meet boundary conditions to high precision:

- Rivers (head- or resistance-specified).
- Lakes (head-, resistance-, or extraction-specified).
- Wells (discharge-specified).
- Recharge/leakage zones.
- Polygonal and circular in-homogeneities in conductivity.
- Thin in-homogeneities (for cut-off walls and fractures).
- Normal-discharge specified boundaries.

More details can be obtained from http://www.groundwater.buffalo.edu/software/VBB/VBBMain.htm.

**Numerical models**

**Saturated flow models**

PLASM, as distributed by the International Ground Water Modelling Centre (IGWMC), is a program package based on the Prickett Lonnquist Aquifer Simulation Program (PLASM) first published by Prickett and Lonnquist (1971). It consists of three finite-difference simulation programs and a pre-processor. The programs simulate 2-D non-steady flow of groundwater in heterogeneous anisotropic aquifers under water table, non-leaky, and leaky confined conditions. Included are options for time-varying pumpage from wells, induced infiltration from streams or shallow aquifers, and water-table-depth-dependent evapotranspiration. The finite-difference equations are solved using a modified alternating direction method. More details can be obtained from http://typhoon.mines.edu/software/igwmcsoft/.

FLOWNET (van Elburg et al. 1993) is used for interactive modelling of two-dimensional steady-state flow in a rectangular, cross-sectional representation of a heterogeneous, anisotropic aquifer. It generates a flow net, composed of flow lines and equipotential lines, obtained by a five-point finite difference approximation to calculate heads and linear interpolation to
determine equipotential lines. The matrix equation is solved using the conjugate gradient
method. The streamlines are determined from the flow function, which in turn is determined
using the adjoint function of the potential function. The model handles no-flow and constant
hydraulic head boundary conditions, variable along the boundary.

More details can be obtained from http://typhoon.mines.edu/software/igwmcssoft/.

**Variably-saturated flow models**

HYDRUS-2D is a finite element program for simulating flow and transport in variably saturated
media (Šimůnek et al. 1999). The unstructured finite element mesh generator is a versatile
tool for better representation of complex geometries. Flow may be simulated in three different
types, vertical flow, axi-symmetric flow, and horizontal flow. The latter option simulates steady
state, saturated flow in a single layered aquifer of any areal shape. The output pressure heads
can be imported into a contouring package to produce a steady state three-dimensional
groundwater table. The ‘vertical flow’ option in HYDRUS-2D models variably saturated flow
with a range of boundary conditions such as a time-varying constant –head boundary, which
simulates a stream. The advantage here is the full flexibility of the code to handle the extent of
stream penetration including the option of modelling an unsaturated zone below the stream. It
is also possible to explicitly model a stream of any shape and it’s full interaction with the
aquifer by assuming it as a material with a porosity equal to unity and having a high hydraulic
conductivity. However, due to the complexity of the code, there are spatial limitations in terms
of the size of the domain that can be modelled; HYDRUS-2D is usually used on a plot-scale
where the flow domain may extend to a maximum of a few hundred metres. The code also
models reactive solute transport, heat flow, root water uptake, and includes an inverse
parameter estimator. The latest version of the code has the ability to simulate groundwater–
surfacewater interactions in constructed wetlands (Langergraber and Šimůnek 2005). More
details can be obtained from http://www.pc-progress.cz.

**Level 3 models; 3-D flow**

**Saturated flow models**

MODFLOW is a modular, finite difference program for simulating three-dimensional saturated
flow. The structure of the code includes a main program and a number of packages that
handle different features of the hydrologic system such as areal recharge, evapotranspiration
or flow to rivers or drains. The interaction of groundwater and surfacewater in MODFLOW is
simulated using either the river or stream packages.

The river package in MODFLOW treats the river as a fixed-head source during a stress
period. The stream aquifer interconnection is represented as a simple conductance through
which one-dimensional flow occurs. It is assumed that measurable head losses between the
stream and the aquifer are limited to those across the streambed layer itself. Conductance is
usually chosen arbitrarily then adjusted during calibration (McDonald and Harbaugh 1988).
Flow between the stream and groundwater system depends on the hydraulic conductance of
the stream aquifer interconnection and the head difference between the stream and the
aquifer. The parameters required for the river package are as follows: river stage, river bottom
elevation, and conductance. The latter is a parameter representing the resistance to flow
between the surfacewater body and the groundwater; it is calculated from the length of a river
reach through a cell \(L\), the width of the river in the cell \(W\), the thickness of the riverbed \(M\),
and vertical hydraulic conductivity of the riverbed material \(K\) and is equal to \(KLM/W\).

The stream package (STR1 Stream Flow Routing Package, Prudic 1989) accounts for the
amount of flow in stream and simulates the interaction between surface streams and
groundwater. Streams are divided into segments and reaches. Each reach corresponds to
individual grid cells, while segments consist of a group of grid cells connected in downstream
order. Stream flow is accounted for by specifying the inflow for the first reach in each segment,
and then calculating the stream flow to adjacent downstream reaches in each segment as
equal to inflow in the upstream reach plus/minus leakage from/to the aquifer in the upstream
reach. The stream flow routing package requires the following parameters: stream inflow,
stream stage, streambed top and bottom elevations, conductance, stream width, and channel’s slope and Manning’s roughness coefficient.

MicroFEM is a program that handles the process of groundwater modelling, from the generation of a finite-element grid through the stages of pre-processing, calculation, post-processing, graphical interpretation and plotting. Confined, semi-confined, phreatic, stratified and leaky multi-aquifer systems can be simulated with a maximum of 20 aquifers. Irregular grids, as typically used by finite-element programs, have several advantages compared to the more or less regular grids used by finite-difference codes. Depending on the type of model this can be extended with layer thicknesses, storativities, spatially varying anisotropy, top system and user-defined parameters. More details can be obtained from http://www.microfem.com.

DYNFLO is a finite element code that simulates fully 3-D multi-layer aquifer systems and allows a wide range of stresses and boundary conditions to be applied. It also has 1D elements for simulating multi-layer wells, under drains, and fractured rock interconnections, and 2D elements, which can represent fault barriers and slurry walls. It can run in steady state or transient mode, and allows for input data updating at any time step during transient runs. DYNFLO has standard fixed head, fixed flux and 3rd type (head dependent flow) boundary conditions, plus special ‘rising water’ and ‘pond element’ features. More details can be obtained from http://www.dynsystem.com/system/dynflow.html.

Variably-saturated flow models

FEFLOW is a software package for modelling fluid flow and transport of dissolved constituents and/or heat transport in the subsurface. The three-dimensional code can handle time-constant or transient boundary conditions. The additional features that FEFLOW has compared to other finite element codes are the ability to simulate variable-density flow and fracture flow. More details can be obtained from http://www.wasy.de/english/produkte/feflow/index.html.

HYDRUS-3D is the 3-dimensional version of HYDRUS-2D.

The MODFLOW-2000 (Harbaugh et al. 2000) version of MODFLOW supports the Variably Saturated Flow (VSF) Process of Thoms et al. (2006). This collection of five packages simulates 3-D variably saturated flow whereby the saturated ground-water flow equation is expanded to include unsaturated flow using Richards’ Equation and solved using a finite-difference approximation.

MODHMS also supports 3-D variably saturated subsurface flow using a Richard’s equation approach.

3.2 Surfacewater models

In this section, a general overview of the most common surfacewater models is given, along with a description of the functionality for representing groundwater–surfacewater interaction.

Level 1; models GW–SW interaction at the conceptual level

AQUATOR (Oxford Scientific Software 2004) is a component-based water resources hydrological model that uses operating rules to represent water resource system management and simulates system dynamics at a daily time-step. The code is based on VBA (Microsoft Visual Basic for Applications), and therefore the components are customizable. AQUATOR has 24 standard components, which are used to represent hydrological elements of the system. The reader is directed to Chapter 2 of the user manual (Oxford Scientific Software 2004) for a complete description of these components, which include links, abstraction, diversion, pump stations, reaches, reservoirs, confluences, catchments amongst others. The code also has a Groundwater component, which is used to represent a simple groundwater source (e.g. river augmentation). The AQUATOR groundwater component does not model groundwater storage, pumping, or recharge in a direct manner. More details can be obtained from http://www.oxscisoft.com/aquator/index.htm.
BIGMOD (MDBC 1996, 2002) is specific to the River Murray system, which is represented in the model as a number of river reaches and branches. The major processes modelled by BIGMOD are:

- The routing of flow and salinity.
- Losses.
- Inflows.
- Extractions.
- The operation of storages and weirs based on specific rules.
- The diversion of water to river branches.

The various flow components of the BIGMOD water balance equation are defined as (MDBC 2002)

\[
q_{\text{out}} = q_{\text{in}} + S_{\text{start}} - S_{\text{end}} - d - \frac{E A}{100} - L_{hf} - L_{cm}
\]  

(4)

where

- \(q_{\text{out}}\) = daily flow out of the reach (ML),
- \(q_{\text{in}}\) = daily flow into the reach (ML),
- \(S_{\text{start}}\) = reach storage at the start of the day (ML),
- \(S_{\text{end}}\) = reach storage at the end of the day (ML),
- \(d\) = daily distributed diversion (ML),
- \(E\) = daily net evaporation rate (mm),
- \(A\) = reach surface area (ha),
- \(L_{hf}\) = daily high flow losses from reach (ML),
- \(L_{cm}\) = daily constant monthly losses (ML).

Other system features that are modelled in BIGMOD include point diversions, drains or tributaries, regulators, weirs, and lakes (off-river storages). The model uses rating tables (flow versus river water level) to calculate river levels, which are used to determine flows into and out of lakes.

At present, the BIGMOD model does not explicitly simulate flow exchanges between groundwater and surfacewater, although the model simulates losses on the floodplain as a function of stream flow (i.e. high flow losses \(L_{hf}\)). In the simulation of salt loads, BIGMOD incorporates groundwater-based salt influxes to the system using 'control variables', which provide the functionality to model complex features (both flow and salinity) and operating rules, without the need to modify the FORTRAN code (MDBC 2002).

REALM (REsource ALlocation Model) is a river hydrology model developed to simulate the operation and optimisation of water supply systems (http://www.dpi.vic.gov.au/DSE/wcmn202.nsf/LinkView/4CE5EDD70A17665ACA257065001B60785CDA9A032D09E6BDCA2570680060E85).

REALM is widely applied in water resources planning and management projects in Victoria, and is supported by the Victorian Department of Sustainability and Environment (DSE) (Victoria University and DSE 2005). REALM adopts a node-link (or node-"carrier") approach to the simulation of river hydrology, and is able to represent reservoirs, demand centres, waterways, pipes, and most other riverine features, plus a wide range of operating rules (Victoria University and DSE, 2005). While the REALM User Manual (Victoria University and DSE, 2005) suggests that the model has a capability for simulating surfacewater-groundwater interaction, although there is little documentation on this aspect of the model.

IQQM (Integrated Quantity and Quality Model) is a hydrological modelling tool that is designed to simulate river system behaviour and the impacts of water resource management policies or changes in policies on stakeholders (Centre for Natural Resources 1999). The capabilities of IQQM are somewhat version-dependent, but in general, the major processes that are simulated include:

- Flow routing in streams, including branches, loops and tributaries.
- Reservoir operation.
- Losses from the system (evaporation, seepage, etc).
- Irrigation.
Review of groundwater–surfacewater interaction modelling approaches

- Off-river storages.
- Fixed demands.
- Floodplain detention storage.
- Groundwater–surfacewater interaction.
- Wetlands.

IQQM represents river systems through links and nodes, and a GUI (Graphical User Interface) is available to assist with model data input and to all the importation of GIS layers. (Catchment Modelling Toolkit 2005a, 2005b, 2005c).

IQQM has been used to simulate groundwater–surfacewater interaction by representing aquifers as off-river storages using ‘traditional’ IQQM functions – i.e. combinations of surfacewater modules rather than a groundwater module (e.g. Water Assessment Group 2005). The approach to modelling groundwater–surfacewater interaction IQQM nodes usually applied to surfacewater features, as surrogates for representing groundwater discharge and recharge. An example of this approach is described in the DNRW IQQM modelling report of the Moreton catchment (Water Assessment Group 2005), in which groundwater–surfacewater interaction has a significant influence on the flow dynamics of Cressbrook Creek and the Lockyer Valley streams. In total, the Moreton IQQM model simulated 29 aquifer units. Flow between stream nodes and adjoining aquifers were modelled using Flow Control Tables (FCT), which specify the groundwater storage-dependent river-aquifer exchange functions. Aquifers were represented as off-stream storages, which fluctuated in storage volume according to bore pumping, rainfall recharge, inter-aquifer exchanges, and river-aquifer exchanges (Water Assessment Group 2005). More details can be obtained from http://mail.toolkit.net.au/mailman/listinfo/iqqm/.

HSPF (Hydrological Simulation Program – FORTRAN) simulates non-point source pollutant loads from urban and rural lands, watershed hydrology and aggregate water quality in natural and man-made water systems. HSPF uses continuous rainfall and other meteorologic records to compute streamflow hydrographs and pollutographs, and incorporates channel, reservoir, constituent and sediment routing effects. HSPF treats the groundwater regime as a lumped entity, and calculates groundwater recharge as the excess from an ‘upper zone’ store. Groundwater discharge is calculated using a base flow recession analogy. More details can be obtained from http://water.usgs.gov/software/hspf.html.

E2 (Argent et al. 2006b) is a node-link model designed for whole-of-catchment modelling of flow and constituent load. E2 represents thecatchment using sub-catchments or ‘functional units’, nodes and links. Functional units may comprise a rainfall-runoff model, a constituent generation model and a filter model. Nodes are used to represent sub-catchment outlets, stream confluences, or other places of interest (e.g. stream gauges). Links serve to store water and to route or process water and constituents passing from node to node, and are used to represent river reaches, floodplains, or dams/storages. Within each link, three models may be assigned - a routing model, a source/sink model and a decay/enrichment model.

The structure of E2 is designed for flexible modelling, where users select from a suite of available algorithms and structures. E2 component models are described by Argent et al. (2006a) and include the rainfall-runoff models AWBM (Australian Water Balance Model), SIMHYD, SURM (Simple Urban Runoff Model), Sacramento and SMAR (Soil Moisture Account Runoff model). E2 also applies the baseflow separation method of Nathan and McMahon (1990) to distinguish between quick and slow flow.

Some of the approaches to rainfall-runoff generation available in E2 consider groundwater pathways, with varying degrees of complexity, although none of the E2 components are designed to simulate regional groundwater flow or the impacts of groundwater management. For example, AWBM adopts a ‘baseflow store’, which is recharged using a fraction of runoff and then depleted at a rate dependent on a base flow recession constant. SIMHYD includes a ‘groundwater store’, which is recharged proportionally to ‘soil wetness’ and discharges to the stream also using a base flow recession constant. SURM also uses a groundwater store approach, and includes both groundwater discharge to surfacewater and groundwater losses to deep seepage. SMAR uses a routing algorithm to simulate groundwater flow attenuation.
and assumes groundwater recharge occurs as an excess of the soil moisture capacity. More details can be obtained from http://www.toolkit.net.au.

The CAT3D framework (PIRVIC, 2005) estimates the impact of various forms of intervention with a combination of paddock/farm scale models and a lateral flow model that are integrated into a regional scale frame work. underlying the surface element network is a 3-dimensional representation of the groundwater system where deep drainage is assigned as recharge to the aquifer. Time-varying groundwater discharge to streams and to the land surface is estimated.

The 2CSalt model (Stenson et al., 2006) is a spatial catchment model for predicting the impact of land-use change on generation of water and salt from upland areas. It was developed as part of the CRC for Catchment Hydrology’s “Catchment Modelling Toolkit”. The modelled area is divided into multiple independent groundwater response units (GRUs), which form the fundamental modelling units. For each GRU, water balance terms (typically obtained from separately run 1D water balance model: PERFECT runs (Littleboy et al., 1992)) are routed through two groundwater stores (hillslope and alluvial), and explicitly assumes a gaining stream. 2CSalt is designed to use broadly available data such as groundwater flow systems (GFS), climate and topography.

SWAT (Soil and Water Assessment Tool) is a watershed model developed by the USDA Agricultural Research Service (ARS). The model was developed to predict land management impacts on water, sediment and agricultural chemical yields (Neitsch et al. 2005). SWAT requires information about weather, soil properties, topography, vegetation and land management practices within the watershed. The catchment is partitioned into sub-basins, which are characterised by relevant information on climate, hydrologic response units, ponds/wetlands, groundwater and the main channel or reach draining the sub-basin. Simulation of the watershed hydrology is separated into two major divisions: the land phase of the hydrologic cycle (i.e. water, sediment, nutrient and pesticide loadings to the main channel), and the water or routing phase of the hydrologic cycle (i.e. the movement of water, sediments, etc) through the channel network to the watershed outlet. More details can be obtained from http://www.brc.tamus.edu/swat/.

**Level 2; explicit modelling of some GW–SW processes**

IQQM Groundwater Module: this module is designed to replicate groundwater storage behaviour in alluvial aquifer systems (Centre for Natural Resources 1999). The IQQM groundwater module was developed in 1999, but has not been in widespread use since its inception, and is not compatible with the current version of IQQM. The module, which is referred to as the ‘IQQM groundwater node’ (specifically IQQM node type 12.1), has been interfaced only with IQQM version 6.31 (Doherty 1999a). The functionality of the IQQM groundwater node is illustrated in Figure 6.

![Figure 6. The functionality of the IQQM groundwater node (following Doherty 1999a).](image)
The IQQM groundwater node was designed specifically to represent near-river alluvial aquifers. The alluvial system is subdivided into an inner and outer alluvium, and flow between the outer boundary (i.e. the Fixed head in Figure 6), the outer alluvium, the inner alluvium and the river is dependent on the respective hydraulic head values that represent each component. Conductance values are used to convert hydraulic head differences to flows. The flows are calculated at each IQQM time step using a simple Euler (i.e. fully explicit) method (Doherty 1999a).

Other than the obvious shortcoming of the restricted application of the method, the IQQM groundwater node also suffers from the following limitations:

- The node requires considerable data input, in terms of system parameterisation (conductance values, aquifer geometry, aquifer storage capacities), and aquifer stresses (e.g. evapotranspiration, recharge, pumping volumes).
- The necessary aquifer parameters and system stresses may be difficult to quantify through field measurements or available data, due to sparsity of observation points, low measurement frequency, and/or the non-physical nature of some of the parameters (e.g. conductance values).
- Some of the model input (e.g. recharge) may require the application of other models – e.g. SWIM (Verburg et al. 1996) or SPLASH (Arunakumaren 1997).
- No attempt at calibrating the node parameters to align model predictions to field observations (e.g. groundwater–surfacewater flows) has been documented.

AQUATOR: The simulation of dynamic river-aquifer interaction (i.e. storage- or discharge-dependent river-aquifer exchange fluxes) is possible in AQUATOR by coupling to HYSIM (Manley and Water Resource Associated 2003; www.watres.com/software/sf-hysim.html). HYSIM (Hydrologic Simulation Model) is a catchment runoff model that simulates rainfall-runoff and accounts for subsurface flow paths using several soil and groundwater stores. River inflow is divided into surface runoff, interflow and base flow. Direct linking of AQUATOR to HYSIM allows the groundwater store of HYSIM to be depleted by the groundwater releases in AQUATOR simulations, thereby reducing the groundwater component of the catchment outflow (i.e. rainfall-driven river inflow) determined by HYSIM. No examples of the application of the AQUATOR-HYSIM model to the simulation of groundwater–surfacewater interaction are evident in the literature.

3.3 Fully coupled groundwater–surfacewater models

Watershed models have been applied to problems of surfacewater management without treating groundwater in much detail. Surfacewater models usually assume that percolation from the soil profile is lost from the system and thus ignore it. Similarly groundwater models have been applied to aquifer management problems without treating surfacewater in any detail. In both cases, the secondary processes (i.e., groundwater processes in surfacewater models or surfacewater processes in groundwater models), are treated either as known inputs or parameters determined by calibration (Sophocleous and Perkins 2000).

In areas with dynamic and hydraulically well-connected groundwater and surfacewater systems, stream aquifer interaction should be simulated using deterministic responses of both systems coupled at the stream-aquifer interface (Swain and Wexler 1996). The need for
integrated surface and groundwater resource management is highly recognized, yet there is still a scarcity of integrated models that can be applied to real world problems. Especially lacking is two-way coupling of watershed and groundwater models, which not only account for inputs to the groundwater system but also for the impact of the groundwater system on the overlying unsaturated zone and, in particular, the root zone (Sophocleous and Perkins 2000). Such interactions are common in areas of shallow water table often encountered near streams and marsh areas.

Most groundwater models are spatially distributed and physically based. They involve numerical solutions of the governing differential equations where the flow domain is discretised into cells, whose sizes depend on system geometry, data availability, and numerical-solution convergence requirements. Watershed models are of the lumped, conceptual type. They have recently been evolving towards the ‘lumped distributed’ models by dividing the basin into topographically based sub-basins with different physical characteristics. Heterogeneity within a sub-basin (arising from different types of soils, vegetation, and land use) is statistically represented by the hydrologic response unit (HRU) concept.

Models that honour groundwater–surfacewater interactions are commonly based on the conductance concept that presumes a distinct interface at the land surface, separating the surface from the subsurface domain (Kollet and Maxwell 2006). Such models link the surface and the subsurface domains via an exchange flux that depends upon the magnitude and direction of the hydraulic gradient across the interface and a proportionality constant (a measure of hydraulic connectivity). This proportionality constant is considered a lumped fitting parameter. The exchange flux appears in both the groundwater and the surfacewater flow equations as general source/sink terms. Recent field studies have shown the absence of a distinct interface between the surface and subsurface, which makes the application of the conductance concept problematic (Kollet and Maxwell 2006).

The holistic approach is the most rigorous where all processes to be modelled are handled directly in a single matrix system that contains all the information about the groundwater–surfacewater model. For example, a lake can be represented in the groundwater model as an unconfined aquifer with a high hydraulic conductivity and a storage coefficient equal to unity. In the decomposed approach the overall system remains partitioned into the natural sub-processes with their own temporal and spatial units and are linked with the appropriate boundary conditions (Kaiser et al. 2000).

An added advantage with coupled modelling over either watershed or groundwater models is improving model calibration by increasing the calibration targets, which reduces parameter non-uniqueness (Sophocleous and Perkins 2000).

The following issues regarding differences in time scales between groundwater and surfacewater processes are worth noting:

- Precipitation and stream flow have much shorter residence time than groundwater.
- Deep drainage and recharge are smaller quantities than precipitation and evapotranspiration, therefore, large time-lumping in evaluating the water balance may mask short periods of recharge by the averaging effect of larger time interval data (Sophocleous 1992).
- Time lumping of soil moisture largely affects water balance components like deep drainage and runoff.
- Large disparities in time steps between surfacewater and groundwater may lead to numerical instabilities. Ideally, both processes should me modelled with the same time step.

The level of complexity is generally determined by the dimensions of the model, as well as the inclusion of all possible physical phenomena in the mathematical interpretation of the coupled system (Gunduz and Aral 2005). In this regard, the most advanced model would involve coupling a three-dimensional surface flow component based on the complete Navier-Stokes equations and a three-dimensional variably saturated subsurface flow component. Due to the large computational powers and high data requirements of such models, modellers choose to reduce model dimensions for large-scale applications (Gunduz and Aral 2005). For example,
Kollet and Maxwell (2006) used a two-dimensional surface flow component model with a three-dimensional variably saturated subsurface flow component. Various researchers have used one-dimensional surface flow component model with either a two-dimensional or three-dimensional subsurface flow component that models variably saturated flow or saturated flow only e.g., (Kollet and Maxwell 2006), (Osman and Bruen 2002), (Swain and Wexler 1996).

Further simplification may involve de-coupling of the surface and groundwater processes; e.g., groundwater response functions can lump groundwater processes and feed them into surfacewater models. Modellers almost always prefer to use less complex models because they are simpler to handle and usually have less data requirements. However, oversimplification may lead to unacceptable errors. For example, Kaiser et al. (2000) showed that to fully account for the filling behaviour of a lake its volume must be carved out of a 3-D groundwater model; intolerable errors occur when using a 2D model.

Level 1; reach-scale lumped conceptual models

The characteristics for a potential level 1 reach-scale model are hereby described. Groundwater–surfacewater interactions within a river reach may be modelled conceptually using a single bucket. We consider a stream reach to which we have inlet and outlet hydrographs from gauging stations. We want to model GW–SW interactions in this reach to account for differences between the inlet ($Q_1$) and outlet ($Q_2$) hydrographs. We conceptualise the problem as follows:

- Represent the system by a single bucket with a boundary that acts as an interface between the stream and groundwater.
- Assume initial steady state conditions with a hydraulic gradient to or from the river (gaining or losing stream, respectively) resulting from some base recharge and AET.
- The magnitude of the gradient and its direction is obtained from field measurements and/or groundwater modelling.
- We use the stage hydrograph as a boundary condition for modelling groundwater processes.
- Discharge to or from the stream (gaining, losing; $\pm Q_s$) varies with hydraulic conductivity, stream conductance, stage height, and head gradient in aquifer (etc.). It is a function of the cumulative effects of base recharge and ET and the above factors.
- During (initial) non-event (base flow) conditions: $Q_1 - Q_2 - \pm Q_s = 0$; we can use this mass balance equation to calibrate $Q_s$ (or any one or combination of the factors that affect it mentioned the above dot point).
- We consider GW–SW processes including bank storage, pumping, evapotranspiration (AET), and flooding and their interactions as follows:
  - Model departures from base AET as a result of groundwater table fluctuations (driven by other processes); note that increased AET when water table is shallow (e.g. during a flood event) will cause a water table drawdown.
  - Model departures from base recharge and its effects on groundwater table levels.
  - Model bank storage dynamically using the stage hydrograph as one continuous time-series boundary condition; the groundwater fluctuations will affect near-field AET.
  - Model near-field and far-field pumping; head drawdowns will affect groundwater AET.
  - Model the frequency and extent of over-bank flooding.

Level 2; catchment-scale river models with groundwater responses

Since the governing groundwater flow equation is linear and time-invariant, linear system theory can be applied via the principle of superposition (Bear 1979), allowing individual excitation’s events to be calculated independently and their responses linearly combined. Linearity in groundwater flow indicates time-independent transmissivity, storage coefficient, and boundary conditions. Pulido-Velazquez et al. (2005) stated that the stream aquifer interaction in any aquifer with linear behaviour, regardless of its heterogeneity, geometry, and
boundary conditions, is equivalent to the summation of the drainage of an infinite number of independent linear reservoirs.

It is advantageous to implement this modelling concept within a GIS modelling framework. The pre-processing and post-processing of data using GIS tools greatly expands the potential use for models as management tools. Shannon et al. (2000) presented prototype software which provides a metadata framework for linking GIS coverages with the procedural MODSIM river basin network flow model, with application to the Lower Snake River flow augmentation.

The groundwater responses can be estimated from analytical solutions (e.g., Knight et al. 2005) or from numerical models such as MODRSP (a modified version of MODFLOW), which calculates kernel or response functions for stream aquifer interaction. MODRSP is an appropriate tool for calculating response functions coefficients since it allows modelling of a multi-aquifer groundwater flow system as a linear system with irregularly shaped areal boundaries and non-homogeneous transmissivity and storativity.

Fredericks et al. (1998) presented a decision support system (DSS) for conjunctive water management of surfacewater and groundwater under prior appropriation; it includes the capability of utilizing groundwater response coefficients generated from a groundwater flow model and a management capability for analysis of various conjunctive use scenarios. The spatially distributed stream-aquifer response coefficients can be used to allocate groundwater return/depletion flows to multiple return/depletion flow node locations anywhere in the basin network. The DSS is constructed around the river basin network flow model MODSIM; it uses stream depletion factors generated from MODRSP. The stream depletion factor is defined as the time where the volume of stream depletion reaches some arbitrary fraction of the net volume pumped during some time period. The stream depletion factor value at any location depends on the integrated effects of irregular impermeable boundaries, stream meanders, aquifer properties, areal variation, distance from the stream, and hydraulic connection between stream and aquifer. Fredericks et al. (1998) also used analytical methods to calculate stream depletion factors and concluded that they result in lower net river return flow values when compared with results from the numerically-based finite difference coefficients.

Zarriello and Ries (2000) developed a precipitation-runoff model for analysis of the effects of water withdrawals on stream flow, and implemented in the Ipswich River Basin, Massachusetts, USA (HSPF STRMDEPL). They used an analytical solution to compute time series of stream flow depletions resulting from ground-water withdrawals at wells. The flow depletions caused by pumping from the wells were summed along with any surface-water withdrawals to calculate the total withdrawal along a stream reach. Model-fit analysis indicated that the simulated flows matched observed flows over a wide range of conditions.

Level 3; catchment-scale coupled GW–SW models

Regardless of the complexity, the coupling process in all models is based on the idea of solving for the common parameters linking the surface and subsurface components (Gunduz and Aral 2005). The solution can be done iteratively by solving both systems individually until the difference between successive solutions becomes less than some prescribed tolerance limit; such models will be referred to as ‘hybrid models’. Alternatively, the surface and the subsurface flow equations can be solved simultaneously within the same global matrix structure; the elimination of the iterative process makes the solution faster (Gunduz and Aral 2005); such models will be referred to as ‘integrated models’.

Hybrid models

A coupled model was developed by combining the USGS models MODFLOW and BRANCH, the interfacing code is referred to as MODBRANCH. BRANCH is a one-dimensional numerical model that simulates unsteady flow in open channel networks by solving the St. Venant equations (Schaffranek et al. 1981). Terms that describe leakage between stream and aquifer as a function of streamed conductance and differences in aquifer and stream stage were added to the continuity equation in BRANCH. Total mass balance in the coupled model is conserved (Swain and Wexler 1996). The BRANCH model calculates new stream stages for each time interval in a transient simulation based on upstream boundary conditions, stream
properties, and initial estimates of aquifer heads. Then, aquifer heads are calculated in MODFLOW based on stream stages calculated by BRANCH, aquifer properties, and stresses. The iterative process continues until convergence criteria for head and stage are met. Due to different time scales encountered in both systems, the model handles multiple BRANCH time intervals within one MODFLOW time step.

Sophocleous and Perkins (2000) presented an intermediate complexity, quasi-distributed, large watershed model, which falls between the fully distributed, physically based hydrological modelling system of the type of the MIKE-SHE model and the lumped, conceptual rainfall-runoff modelling system of the type of the Stanford Watershed Model (Crawford and Linsley 1966). This is achieved by integrating the quasi-distributed watershed model SWAT with the fully distributed groundwater model MODFLOW. The advantages with the approach is the smaller data requirements, the statistical handling of watershed heterogeneity by employing the hydrologic-response unit concept, and the flexibility in handling stream aquifer interactions, distributed well withdrawals, and multiple land uses.

Jagelke and Barthel (2005) presented a study dealing with groundwater modelling within the framework of the integrated regional model MOSDEW, which is being developed as part of the EU Water Directive. The groundwater model interacts with the rainfall-runoff model HBV and with the water-demand model WEAP. The groundwater model receives the spatially and temporally highly differentiated groundwater recharge from the HBV-model and water extractions on municipality scale from the WEAP-model. Using the data as the boundary conditions, the groundwater model simulates the base flow and provides it to the flood routing module of the HBV-model.

Lin and Medina Jr (2003) incorporated the transient storage concept in a conjunctive stream aquifer model. Three USGS models were coupled together: (1) MODFLOW handles groundwater flow in the aquifer; (2) DAFLOW computes unsteady stream flow by means of the diffusive wave routing technique, as well as stream-aquifer exchange simulated as streamed leakage, and (3) MOC3D compute solute transport in the groundwater zone. The conjunctive stream aquifer model with transient storage can handle well the bank storage effect under a flooding event. The stream aquifer interaction is such a strong sink/source for solute transport in streams that it must not be ignored in simulation.

Osman and Bruen (2002) studied seepage from a stream, which partially penetrates an unconfined aquifer, for the case when the water table falls below the streambed level. They presented a simple method that considers the effect on seepage flow of suction in the unsaturated part of the aquifer below a disconnected stream and allows for the variation of seepage with water table fluctuation. The technique is incorporated into MODFLOW and tested by comparing its predictions with the variably saturated code SWMS-2D. It is shown that the current approaches underestimate the seepage and associated local water table mounding, sometime substantially.

Although GFLOW is an analytic element, conjunctive surfacewater and groundwater flow model based on the Dupuit-Forcheimer assumption, but it does support three-dimensional streamline tracing. The program includes features for simulation of line sinks and sources, injection and withdrawal wells, ponds, regional uniform flow, and areal recharge. The sources and sinks can be specified by recharge/discharge rates or by heads. Locally, transient flow due to a well or fully three-dimensional features may be included (e.g. partially penetrating well, shallow pond). Furthermore, the model can handle local heterogeneity and anisotropy. Feinstein et al. (2006) have recently coupled GFLOW to MODFLOW2000 where the former is used to represent the water table layer as it better handles groundwater–surfacewater interactions, routing of stream flow, and exact location of surface-water and well elements than does MODFLOW. By inserting a truncated MODFLOW model under a modified version of GFLOW, it is possible to take advantage of the flexibility of analytic elements in simulating local conditions while still preserving the three-dimensional flow field and the interplay of shallow and deep stresses.
Review of groundwater–surfacewater interaction modelling approaches

Fully integrated models

Kollet and Maxwell (2006) presented a coupled model that incorporates a two-dimensional overland flow simulator into the parallel three-dimensional variably saturated subsurface flow code ParFlow; the overland flow simulator takes the form of an upper boundary condition and is, thus, fully integrated without relying on the conductance concept. Applying the model shows the propagation of uncertainty due to subsurface heterogeneity to the overland flow predictions.

Monninkhoff (2002) coupled the WASY groundwater software FEFLOW with the DHI surfacewater software MIKE11. The coupling is not iterative. After each time step, discharges calculated by FEFLOW to the coupled boundary points are exported to MIKE11 as an additional boundary condition. MIKE 11 calculates its time step as often as needed to reach the actual time level of FEFLOW. The actual water levels in MIKE11 are then exported to the FEFLOW coupling boundary nodes and the time stepping continues.

GSFLOW (Niswonger et al. 2006) is a new USGS model for groundwater–surfacewater interactions. GSFLOW couples PRMS (Precipitation Runoff modelling System) to MODFLOW with a new family of modules for simulating processes in the unsaturated zone; it includes enhancements to both PRMS and MODFLOW to facilitate their dynamic coupling. GSFLOW uses physically-based equations to describe critical processes in the soil zone (the uppermost part of the unsaturated zone), including infiltration, runoff generation, and lateral flow in temporarily saturated material. Flow in the unsaturated zone is based on a 1-day kinematic-wave approximation to the Richard's equation, implicitly coupled to MODFLOW.

MODHMS (HydroGeologic-Inc. 2001) is a MODFLOW-based simulator for fully integrated groundwater–surfacewater and water quality modelling. MODHMS includes dynamic interactions between overland flow, channel flow, and groundwater to simulate water supply management scenarios, flood control and river flow analyses, and wetland restoration analyses. Additional modules have been incorporated into the popular MODFLOW code to provide a physically-based, spatially-distributed conjunctive surface/subsurface modelling framework that includes:

- 3-D variably saturated subsurface flow (Richards equation).
- 2-D aerial overland flow (diffusion wave approximation).
- Flow through a network of 1-D channels or pipes (diffusion wave approximation, with Priesmann Slot conceptualization for pressurized flow in pipes).

The hydrologic cycle is represented as a fully integrated system with dynamic interactions between all regimes of flow. MODHMS can accommodate a variety of simulation scales from regional to local, i.e. from several thousand square km to less than one ha. Numerous surface and subsurface features are incorporated into MODHMS for rural and urban watersheds, including hydraulic structures such as dams and weirs. Comprehensive interception and evapotranspiration calculations are also included, with various potential evapotranspiration formulas (ET from bare ground or vegetated surfaces), which may be applied in different regions of a model domain. Analytic infiltration formulas may also be applied in regions of a domain where the subsurface is not explicitly modelled. Hydraulic structures, withdrawals, and flow regulation schemes are also incorporated to provide a tool for analysis as well as management of water budgets. (Bedekar et al. 2006) used MODHMS to analyse the operations of the marsh driven operation plan for pumps and detention basins.

MIKE SHE (DHI-Water-and-Environment 2005) is an advanced integrated hydrological modelling system. It simulates water flow in the entire land based phase of the hydrological cycle from rainfall to river flow, via various flow processes such as, overland flow, infiltration into soils, evapotranspiration from vegetation, and groundwater flow. Fully dynamic exchange of water between all major hydrological components is included, e.g. surfacewater, soil water and groundwater (integrated). It solves basic equations governing the major flow processes within the study area (physically based). The spatial and temporal variation of meteorological, hydrological, geological and hydrogeological data across the model area is described in gridded form for the input as well as the output from the model (fully distributed). MIKE SHE is by default coupled to MIKE 11. MIKE-SHE has two options for coupling surfacewater and
groundwater representations: river-aquifer exchange and area-inundation flooding (DHI 2005). Both these methods estimate head-dependent exchange fluxes, with river-aquifer exchange being considered a line source/sink process and area-inundation flooding being simulated as the infiltration through either unsaturated or saturated subsurface conditions of ponded water over the flooded area. More details can be obtained from http://www.dhisoftware.com/mikeshe/.

HydroGeoSphere (Therrien et al. 2006) is based on Frac3DVS (Therrien and Sudicky 1996), but has additional capabilities to simulate fully integrated surface-subsurface flow and transport, in addition to variable saturation and density-dependent flow processes. It is a fully distributed model with simulation capabilities that span the terrestrial hydrologic cycle. A feature of HydroGeoSphere is that rainfall inputs are allowed to partition into components such as overland and stream flow, evaporation, infiltration, recharge and subsurface discharge into surfacewater features such as lakes and streams. Surfacewater flow is calculated by using the two-dimensional diffusive wave approximation of the St Venant equations. Flow through the unsaturated zone is approximated by using the Richards equation and several options exist for defining relationships between saturation, pressure and hydraulic conductivities. HydroGeoSphere allows for the discretisation of the problem domain using either finite-difference or finite-element approaches. The numerical implementation is based on the robust Newton-Raphson linearization method.
4 Discussion and summary

The term ‘horses for courses’ is commonly used in the modelling world; it refers to how critical model choice is. In order to choose the right modelling tool, we need to have conceptualised our problem correctly, this implicitly means that we know a priori the processes involved and how they interact with each other. The next step is then choosing (or developing) a modelling tool with the ability to model these processes at the required spatial and temporal scales. The latter is closely related to the processes themselves; for example, surfacewater processes are fast whereas groundwater processes have a much slower response time. Spatial scales are dictated by the questions that are being posed, i.e. are we interested in the response of a whole catchment, a river reach, or a few metres of riparian land in the vicinity of a stream. The spatial scale will dictate the level of model complexity where larger scale models usually adopt a conceptual approach and smaller scale models adopt a more physically based approach. This affects data requirements with the latter usually requiring much more data than the former. The landscape setting where the model is being applied also has a significant effect on model choice. For example, modelling water flow in fractured media require dual-porosity models; layered systems require models that can handle heterogeneity and landscapes with large flat floodplains need a good modelling engine for handling evapotranspiration and overbank flooding.

The suitability of models to address the research questions posed in this project is assessed against a set of criteria that were deemed necessary if these models were to be successfully applied to the relevant Australian landscape settings at the correct spatial and temporal scales. The deficiencies of current models as identified in this study are related to the following issues.

- **GW–SW interaction processes.** These processes are poorly handled in existing surfacewater models and groundwater models. In river models, this interaction is treated simply as a loss term. In groundwater models, the river is simplistically modelled as a boundary condition. More sophisticated models that explicitly account for the GW–SW interaction usually require more data, which is not always readily available. Furthermore, such models require a very high degree of modelling expertise, which is not always available in water management agencies.

- **Identifying the GW–SW interaction processes that are most relevant to the Australian landscape is very critical.** When choosing modelling tools for groundwater–surfacewater interaction, it is important to strike the right balance between surfacewater processes and groundwater processes. This balance can only be achieved when special-purpose in-house models are developed for the right purpose.

- **Data availability and levels of complexity.** In many cases, model choice is restricted by data availability. Data requirements are closely related to model complexity and the spatial scale at which the model operate. The scarcity of field data is highly recognised. High fidelity complex models that may be readily available require intensive field measurements to support it. At a whole-of-river scale, readily available data can only support low fidelity modelling. Therefore, we need to develop simple models that require less data and are tailored for the Australian landscape.

- **Temporal and spatial scale issues.** The spatial scale at which a model is applied dictates its level of complexity and hence what processes are, and aren’t accounted for. Large-scale models usually adopt a lumped approach that requires less parameterisation whereas smaller scale physically based models can explicitly account for more processes. We need to develop models that evaluate GW–SW interaction that are compatible with the river models currently used in Australia (such as IQQM); that is, the spatial scale of the developed model needs to be compatible with the river model node spacing. The issue of temporal scales becomes critical when modelling groundwater–surfacewater interactions because surfacewater processes are quick whereas groundwater processes are much more attenuated. Large time lumping in evaluating a particular process may mask other
processes that may occur during short periods as a result of the averaging effect of larger time interval data. Custom-built models within these projects would ensure that no such disparities in time steps occur and that time stepping is consistent with the models currently used in Australia.

- Landscape settings. The developed models would be consistent with the main landscape settings that were identified in Reid et al. (2008), namely: (1) upland fractured rock systems; (2) layered fractured rock systems e.g. sandstones, basalts; (3) contained alluvial valleys; and (4) regional systems.

- Availability and GIS modelling environment. Another issue that affects model suitability for the purposes of the D3/AHMI-GSWIT projects is whether or not software packages are freely available or not. The source code of proprietary software is not readily available for further development. This issue is highly relevant for the GIS modelling framework through which the models will be delivered. The National Water Commission has stipulated that the developed tools should be made freely available to all Australian water users, hence models that use proprietary GIS software are deemed useless. The proposed modelling tools will be developed using the in-house TIME environment of the CRC Catchment Hydrology. More sophisticated models such as MODFLOW, which can be used to test the validity of the proposed models, should also have interface protocols developed to ensure that can be used with existing and new TIME-based river models.

Tables 2 and 3 categorise some of the models covered in this review in terms of their capacity to model the most critical processes relevant to surfacewater groundwater interactions. This capacity is presented in a hierarchical manner. As discussed earlier, there are a number of criteria that we need to be considered when choosing a tool to model a particular problem. The criteria may be generic such as scales or project-specific such as the capacity to develop in TIME as in the case of the D3/AHMI-GSWIT projects. Referring to Table 4, we look at the suitability of some models from a process point of view. Analytical solutions (the Theis type) have been developed to solve stream depletion problems; they can also be used in a reverse manner to model groundwater head and flux responses to a stream with stage fluctuations (bank storage) but have not been developed to model evapotranspiration. MODFLOW, being the most commonly used tool to model groundwater flow using a sophisticated finite difference method has a simplistic model for evapotranspiration. The HYDRUS family of models has the capacity to handle all five processes at the highest level of complexity. However the model operates at a relatively small scale, has high data requirements, is a proprietary product, and cannot be implemented in TIME. Hence, we conclude that such a package cannot be adopted to address the needs of the D3/AHMI-GSWIT projects but might be useful for modelling a small domain where model parameters and calibration data is abundant; the outcomes of such modelling experiments enhances our understanding of the system. In contrast, analytical solutions that have less data requirements can be further developed and easily implemented in TIME and hence are favourable potential candidates as modelling tools for the D3/AHMI-GSWIT projects.
Table 2. Process-based categories for groundwater models. 0 – does not model the process; 1 – simple lumped approach; 2 – lumped, spatially explicit; 3 – fully distributed, process-based.

<table>
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<th>Process-based category</th>
<th>Bank storage</th>
<th>Stream depletion</th>
<th>AET</th>
<th>Wetland interaction</th>
<th>Flooding</th>
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<td>2</td>
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</tr>
<tr>
<td>FEFLOW-MIKE11</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>MODHMS</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>MIKE-SHE</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>HydroGeoSphere</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>3</td>
</tr>
</tbody>
</table>

Table 3. Process-based categories for surfacewater models. 0: does not model the process; 1: simple lumped approach; 2: lumped, spatially explicit; 3: fully distributed, process-based; 4: some capability through inter-model linking.

<table>
<thead>
<tr>
<th>Riverine process</th>
<th>AQUATOR</th>
<th>BIGMOD</th>
<th>E2</th>
<th>HSPF</th>
<th>IOWO</th>
<th>MIKE-SHE</th>
<th>REALM</th>
<th>SWAT</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. GW–SW interaction</td>
<td>4</td>
<td>0</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>3</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>2. Overland flow and throughflow</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>1</td>
<td>3</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>3. Flow attenuation</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>3</td>
<td>2</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>4. Storages and reservoir operation</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>5. Off-stream storages and wetlands</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>3</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>6. Floodplain processes</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>3</td>
<td>2</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>7. Bank storage effects</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>1</td>
<td>0</td>
<td>3</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>8. Evapotranspiration</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>1</td>
<td>2</td>
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<tr>
<td>9. Point sources/sinks</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>3</td>
<td>2</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>10. In-stream evaporation and rainfall</td>
<td>1</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>1</td>
<td>1</td>
</tr>
</tbody>
</table>
### Table 4. Assessment of various models against the D3/AHMI-GSWIT projects criteria.

<table>
<thead>
<tr>
<th>Model</th>
<th>Dimensions</th>
<th>Heterogeneity</th>
<th>Spatial scale</th>
<th>Data requirements</th>
<th>Availability</th>
<th>TIME</th>
<th>GW/SW processes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Analytical solutions</td>
<td>1D</td>
<td>Possible</td>
<td>Reach scale</td>
<td>Low</td>
<td>Y</td>
<td>Y</td>
<td>GW</td>
</tr>
<tr>
<td>FLOWNET</td>
<td>2D</td>
<td>SS, heterogeneity, anisotropic</td>
<td>Reach scale</td>
<td>Low</td>
<td>Proprietary</td>
<td>N</td>
<td>GW</td>
</tr>
<tr>
<td>CAPZONE</td>
<td>2D</td>
<td>No</td>
<td>Reach scale</td>
<td>Low</td>
<td>Proprietary</td>
<td>N</td>
<td>GW</td>
</tr>
<tr>
<td>GFLOW</td>
<td>2D</td>
<td>SS, single aquifer, heterogeneity.</td>
<td>Reach scale</td>
<td>Low</td>
<td>Proprietary</td>
<td>N</td>
<td>GW</td>
</tr>
<tr>
<td>WinFlow</td>
<td>2D</td>
<td>Transient</td>
<td>Sub-reach scale</td>
<td>Low</td>
<td>Proprietary</td>
<td>N</td>
<td>GW</td>
</tr>
<tr>
<td>HYDRUS-2D</td>
<td>2/3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Up to sub-reach scale</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW</td>
</tr>
<tr>
<td>MODFLOW</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Regional</td>
<td>Medium</td>
<td>Public domain</td>
<td>Y</td>
<td>GW</td>
</tr>
<tr>
<td>FEFLOW</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Sub-reach scale</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW</td>
</tr>
<tr>
<td>MODBRANCH</td>
<td>3D GW 1D SW</td>
<td></td>
<td>Medium</td>
<td>Research tool</td>
<td>GW/SW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SWAT/MODFLOW</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Regional</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW/SW</td>
</tr>
<tr>
<td>FLOW/MIKE11</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Up to reach scale</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW/SW</td>
</tr>
<tr>
<td>GSFLOW</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Regional</td>
<td>Medium</td>
<td>Public domain</td>
<td>N</td>
<td>GW/SW</td>
</tr>
<tr>
<td>MODHMS</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Regional</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW/SW</td>
</tr>
<tr>
<td>MIKE SHE</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Regional</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW/SW</td>
</tr>
<tr>
<td>HydroGeoSphere</td>
<td>3D</td>
<td>Transient/heterogeneity /multiple layer</td>
<td>Regional</td>
<td>High</td>
<td>Proprietary</td>
<td>N</td>
<td>GW/SW</td>
</tr>
</tbody>
</table>
5 Recommendations

The modelling tools that are going to be developed in the D3/AHMI-GSWIT projects aim to estimate exchange fluxes between groundwater and surface-water for lowland rivers and how these may change with groundwater and surfacewater management. The modelling will mainly address issues related to integrated groundwater–surfacewater accounting and groundwater dependent ecosystems. Given the complexity of the GW–SW processes involved and the scarcity of data required to validate and/or calibrate complex models, a simple modelling approach will hence be adopted.

Referring to the complexity levels of modelling approaches outlined above, the review has identified the need to develop Levels 1 and 2 modelling tools suitable for reach and sub-reach river scales that have the capacity to model the GW–SW processes as follows:

- Level 1 complexity, reach scale, ‘GW–SW Link’ model, which operates as a groundwater link to river models. The expected outcome of this model is accounting for GW–SW interactions at the river-reach scale. This will enhance the performance of river models by accounting for the effects of GW–SW interactions that are likely to take place along a river reach. The scale at which this model will operate is envisaged to be in the order of tens of kilometres and should conform to the node spacing of the river model to which it would be coupled. This larger reach-scale tool will model some or all of the groundwater processes but in a simpler lumped manner (but must include far field processes such as recharge, evapotranspiration and far field pumping). A general conceptualisation of this model (for a gaining stream) is shown in Figure 7, which is similar to the conceptualisation proposed by Doherty (1998, 1999a).

![Figure 7. GW–SW interactions for a gaining stream system to be modelled in the Level 1 GW–SW Link model.](image)

- Level 2 complexity, sub-reach scale, ‘Floodplain Processes’ model, which dynamically models bank storage, evapotranspiration, and floodplain inundation. The expected outcome is modelling the GW–SW interactions at the sub-river-reach scale with higher resolution and the capacity to link to ecological response models. The Floodplain Processes model will operate at a scale smaller than that of the GW–SW Link model. It is envisaged that it will be a stand-alone tool to explicitly model floodplain processes on fine spatial and temporal scales; it can also be incorporated into a large scale river model. It aims to simulate, at a high resolution, multiple floodplain processes such as bank storage and floodplain inundation in addition to the processes previously outlined in the GW–SW Link model. The groundwater link to the Floodplain Processes model may be similar to that used in the GW–SW link model. The interaction between various GW–SW processes will be explicitly modelled. Conceptualisation of the relevant GW–SW processes that would be considered in the Floodplain Processes model is shown in Figure 8.
The default data sets for the models considered for application in the D3/AHMI-GSWIT projects need to be clearly identified with a thorough understanding of their sensitivities. This information is vital for informing the experimental design. The parameters that would be required for the various modelling activities are identified to ensure that a fieldwork plan is designed to inform the modelling activities. The various model parameters include:

**GW–SW Link model**
- Riverbed conductance
- Elevation of base of riverbed and alluvium
- Conductance of alluvial-GW interface
- Heads in river and GW
- River flow/head relationship
- Alluvial storage capacity and specific yield
- AET (vegetation type and extinction depth)
- Total pumping volumes

**Floodplain Processes model**
- Aquifer diffusivity; aquifer thickness, specific yield and hydraulic conductivity
- River bed conductance
- Levels of river and aquifer bases
- Pumping schedules
- Pump locations
- River flows and stage height time series
- DEM with stream network
- Groundwater salinity
- Recharge rate
- Tree health-flow
- Vegetation type (for AET)
- Groundwater depth
- Backwater curves

After establishing surfacewater groundwater connectivity at the large scale, the developed models can be linked to river models and thus account for GW–SW interactions at the catchment scale (via a link to the eWater CRC/National Water Commission Projects P2/AHMI-RSMT (Australian Hydrological Modelling Initiative – River Systems Modelling Tool)). Baseline mapping of the degree and nature of GW–SW connectivity at the catchment level is likely to occur as part of the National Water Commission project Mapping Potential Surfacewater/Groundwater Connectivity across Australia. This is a critical data layer for incorporating the developed modules into catchment scale models. Such models will guide water allocation and trade, particularly in the context of salinity credits, integrated GW–SW accounting and the maintenance of minimum stream flows.
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Review of groundwater–surfacewater interaction modelling approaches


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