Journal of Hydrology 538 (2016) 363–373

Contents lists available at ScienceDirect

Journal of Hydrology

journal homepage: www.elsevier.com/locate/jhydrol

Solute transport processes in flow-event-driven stream–aquifer interaction

Yueqing Xie *, Peter G. Cook, Craig T. Simmons

National Centre for Groundwater Research and Training, School of the Environment, Flinders University, Adelaide, South Australia, Australia

ARTICLE INFO

Article history:
Received 23 February 2016
Received in revised form 11 April 2016
Accepted 13 April 2016
Available online 22 April 2016
This manuscript was handled by Peter K. Kitanidis, Editor-in-Chief, with the assistance of Hund-Der Yeh, Associate Editor

Keywords:
Surface water–groundwater interaction
3D river–aquifer modelling
Solute transport
Flow events
Hydrograph separation

SUMMARY

The interaction between streams and groundwater controls key features of the stream hydrograph and chemograph. Since surface runoff is usually less saline than groundwater, flow events are usually accompanied by declines in stream salinity. In this paper, we use numerical modelling to show that, at any particular monitoring location: (i) the increase in stream stage associated with a flow event will precede the decrease in solute concentration (arrival time lag for solutes); and (ii) the decrease in stream stage following the flow peak will usually precede the subsequent return (increase) in solute concentration (return time lag). Both arrival time lag and return time lag increase with increasing wave duration. However, arrival time lag decreases with increasing wave amplitude, whereas return time lag increases. Furthermore, while arrival time lag is most sensitive to parameters that control river velocity (channel roughness and stream slope), return time lag is most sensitive to groundwater parameters (aquifer hydraulic conductivity, recharge rate, and dispersivity). Additionally, the absolute magnitude of the decrease in river concentration is sensitive to both river and groundwater parameters. Our simulations also show that in-stream mixing is dominated by wave propagation and bank storage processes, and in-stream dispersion has a relatively minor effect on solute concentrations. This has important implications for spreading of contaminants released to streams. Our work also demonstrates that a high contribution of pre-event water (or groundwater) within the flow hydrograph can be caused by the combination of in-stream and bank storage exchange processes, and does not require transport of pre-event water through the catchment.

1. Introduction

The interaction between streams and groundwater has been a recent focus of hydrogeologic research (Sophocleous, 2002; Fleckenstein et al., 2010). Groundwater discharge sustains streams during periods of low rainfall, and infiltration from losing streams replenishes groundwater resources. However, exchange between surface water and groundwater is very difficult to measure, particularly at the scales of interest to water managers (Kalbus et al., 2006). A number of studies have therefore used groundwater chemistry to provide regional scale estimates of groundwater discharge to streams – based on the principle that the chemical composition of stream water provides information on processes occurring over the entire water flow path. However, the use of water chemistry to infer groundwater discharge to streams, has usually either: (i) focussed on sampling during stable flow conditions (e.g., Cook, 2013; Harrington et al., 2014; McCallum et al., 2010; Unland et al., 2013), or (ii) applied simplified end-member mixing models to single flow events in small upland catchments (e.g., Buttle and Peters, 1997; Pinder and Jones, 1969). There have not been any studies that have measured spatial and temporal changes in stream chemistry over regional scales, and used this information to infer spatial and temporal patterns of groundwater discharge. But conjunctive management of surface water and groundwater would seem to require such an analysis. Many surface water gauging stations monitor electrical conductivity as well as stream stage (Fig. 1). Because groundwater is usually more saline than stream water, these electrical conductivity records should contain information on groundwater – stream interaction over large spatial and temporal scales, but the data does not appear to be widely used. It is likely that interpreting such data in terms of groundwater discharge to streams requires a better understanding of stream – aquifer exchange processes than we currently possess.

The hydrodynamics of event-driven stream–aquifer interaction has been studied extensively (e.g., Chen and Chen, 2003; Claxton et al., 2003; Squillace, 1996; Whiting and Pomeranets, 1997; McCallum et al., 2010). It is known that during stream stage

* Corresponding author. Tel.: +61 8 82015140.
E-mail address: yueqing.xie@flinders.edu.au (Y. Xie).

http://dx.doi.org/10.1016/j.jhydrol.2016.04.031
0022-1694/

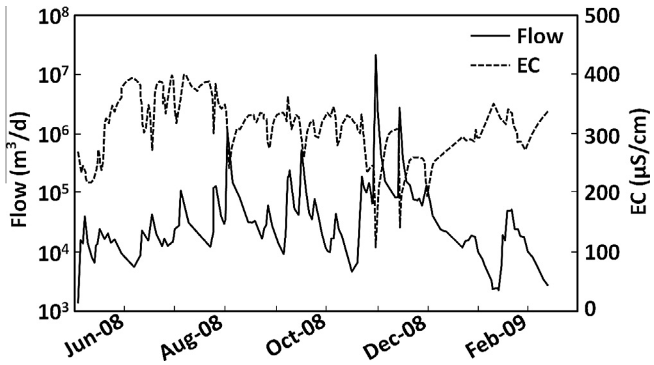

Fig. 1. A field example of time lag between hydrograph and chemograph (after McCallum et al., 2010). Low EC values are associated with peaks in stream flow.

increase, flow event water is stored temporally in both banks and underneath the stream channel (Chen and Chen, 2003; Squillace, 1996). This bank storage water will be released during flood recession and the release time may take many days to several years depending on aquifer hydraulic conductivity (Whiting and Pomeranets, 1997; McCallum et al., 2010; Doble et al., 2012). As the flow pulse from a rain event continues downstream, it will be progressively attenuated by this bank storage (Pinder and Sauer, 1971) and consequently the bank storage exchange will weaken with distance downstream.

In comparison, the solute dynamics of stream–aquifer interaction during flow events has been studied only to a limited extent. McCallum et al. (2010) performed both groundwater flow and solute transport simulations during stream stage fluctuation to investigate chemical baseflow separation, and Gu et al. (2012) simulated reactive transport to examine the effect of bank storage on nutrient and contaminant transport. Although McCallum et al. (2010) and Gu et al. (2012) provided significant insights into particular problems, both studies assumed that the solute concentration of water entering the adjacent aquifers during stream stage rises was constant in time. This assumption is problematic as a number of studies have shown that stream solute concentration changes during flow events (Glover and Johnson, 1974; Walling and Foster, 1975; Kurtenbach et al., 2006). Glover and Johnson (1974) and Walling and Foster (1975) observed that most ion concentrations (e.g., Ca^{2+}, HCO_3^-, Na^+) decrease during stream stage rise and increase back to the original concentration during flow recession. They also observed that the concentration changes are delayed relative to stream stage changes. Kurtenbach et al. (2006) confirmed the time lag between concentration change and stream stage change and found that this time lag can be observed within a short distance (80 m) after a flow pulse is generated. However, the studies of Glover and Johnson (1974), Walling and Foster (1975) and Kurtenbach et al. (2006) do not provide any insights into how the time lag will vary between different flow events.

Thus, although a number of papers have described solute transport during stream flow events, the spatial and temporal variations in stream chemistry that are likely to occur during different flow events in different types of catchments are still poorly understood. Previous studies have modelled 2D transects perpendicular to a stream to simulate solute flux to streams (McCallum et al., 2010; Gu et al., 2012). However, to examine trends in concentrations within the stream requires a 3D model that also simulates flow and transport within the stream channel. The aim of this paper is to provide a systematic and quantitative examination of the transport of conservative solutes during flow events. To do this, we simulate flow events through a highly simplified, synthetic 3D stream–aquifer system and observe the resulting variation in stream water solute concentration. This results in a sensitivity analysis that shows how variations in some important stream and aquifer conditions affect stream solute concentrations. Three characteristics were examined: (1) time lag between the arrival of wave front and solute front at a downstream location; (2) minimum concentration of chemograph; (3) time lag between the recovery of stream stage and stream concentration. Although we focus on conservative solutes in order to simplify the complex 3D model, some solutes may undergo biological, chemical and physical processes, which will be superimposed on the processes described in this paper. The ultimate goal is to be able to infer stream and aquifer characteristics from spatial and temporal changes in stream solute concentrations, and this paper provides a small step towards this goal.

2. Numerical modelling

A synthetic 3D numerical model representing a rectangular aquifer bisected by a stream was established to investigate spatial and temporal variations in solute concentration in streams under various flow events. To simulate stream and groundwater flow and solute transport, the fully coupled numerical code HydroGeoSphere (Therrien et al., 2010) was selected. HydroGeoSphere uses the 1-D diffusion-wave approximation of the Saint Venant equations to simulate surface water flow and uses the 3-D Richards equation to simulate variably saturated groundwater flow. Solute transport in both surface water and groundwater is realised through the advection–dispersion equation. The reader is referred to Therrien et al. (2010) for a detailed description of numerical implementation.

The characteristics of the hypothetical model are loosely based on those of shallow alluvial rivers in semi-arid environments (e.g., Cook et al., 2006). In these systems, runoff is principally generated in steep headwater catchments, where the river is incised directly into bedrock. Our model represents the lowland alluvial section of these systems, where the stream flows across an alluvial plain, and direct runoff to the stream is rare. The length of the model is 20 km and the width is 4 km (Fig. 2). The model thickness increases from 10 m near the stream channel to 12 m at lateral boundaries. The whole domain is slightly inclined at a gradient of 0.001 to allow free flow of surface water. An incised straight stream channel is located in the middle of the model domain with a depth of 1.0 m measured from the top of the stream bank. The stream cross-section is trapezoid-shaped with bottom 10 m wide and top 20 m wide. Due to this symmetric setup, the groundwater divide occurs under the middle of the stream. As a result, we simulated half of the domain with a half stream to reduce computational burden in using the 3-D Richards equation. In this integrated model, only the streambed and the stream bank are treated as surface domain, and the dual node approach with a coupling length of 0.01 m is used.

No-flow boundary conditions were assigned to all side boundaries (apart from the stream itself) and therefore groundwater had to discharge into the stream. A uniform diffuse recharge rate of 0.0001 m/d (i.e., 36.5 mm/y) replenishes the aquifer and maintains stream flow. Because no surface domain was specified at the top of the aquifer and we did not simulate precipitation, there was no overland flow in any simulations. At the inlet of the channel, a no-flow boundary condition was assigned when there was no flow event. Otherwise, a time-varying head boundary condition implemented as a stream water depth $d_{in}(t)$ was imposed to represent incoming flow events generated in the headwater areas. $d_{in}(t)$ was converted to a time-varying stream flow rate using Manning's equation. Waves are symmetric and cosine-shaped, as in several previous studies (Chen and Chen, 2003; McCallum et al., 2010). All waves were initiated at the channel inlet after 0.5 d. The wave equations are thus given by