Supplement A: Mathematical formulation of Groundwater-Surface water interaction model

At each layer $(m = 1 \dots M)$ of an unconfined aquifer, the exact 3D solution to the saturated steady flow governing equation, with no-flow conditions along the sides of the domain, was obtained in terms of discharge potential function $(\phi_m = K_m h_s)$ as (Ameli and Craig, 2014):

$$\phi_m(x,y,z) = \sum_{j=0}^{J-1}\sum_{n=0}^{N-1} cos\, \omega_j x\, cos\, \omega_n y\, [A_{jn}^m cosh(\gamma_{jn} z) + B_{jn}^m sinh(\gamma_{jn} z)] \tag{S.1}$$

where $\omega_j = \frac{j\pi}{L_x}$; $\omega_n = \frac{n\pi}{L_y}$; $\gamma_{jn} = \pi\sqrt{\frac{j^2}{{L_x}^2} + \frac{n^2}{{L_y}^2}}$ for j = 0 ….$J-1$ & n = 0 ….$N-1$

In eq. (S.1), $h_s(x,y,z)$ is the total hydraulic head, and K_m (LT^{-1}) is the m^{th} layer saturated hydraulic conductivity. A continuous map of hydraulic head was then obtained as:

$$h_s(x,y,z) = \frac{\phi_m(x,y,z)}{K_m} \tag{S.2}$$

To complete the solution, the unknown series solution coefficients of each layer (A_{jn}^m, B_{jn}^m) were calculated by imposing infiltration rate along the topographic surface, the no-flow condition along the bottom boundary, and the continuity of flux and head conditions along the layer interfaces of the multi-layer unconfined aquifer. A simple numerical least square scheme was used to impose these boundary and continuity conditions. In general, this continuous solution (Eq. (S.1)) exactly satisfies the mass balance and Darcy equations in the entire watershed, except along the boundaries where mass balance and Darcy equations are prone to numerical least square error. Ameli and Craig (2014) showed that this error can be negligible when sufficient number of control points was used within numerical least square algorithm.

The continuous maps of Darcy fluxes at the m^{th} layer and at each x, y and z directions can be computed by the following equation:

$$q_m^x\,(x,y,z) = \frac{d\phi_m\,(x,y,z)}{dx} \tag{S.3a}$$

$$q_m^y\,(x,y,z) = \frac{d\phi_m\,(x,y,z)}{dy} \tag{S.3b}$$

$$q_m^z\,(x,y,z) = \frac{d\phi_m\,(x,y,z)}{dz} \tag{S.3c}$$

Equation (S.3) can also be used to determine groundwater discharge fluxes at discharge areas (seepage faces) along the land surface as well as groundwater recharge fluxes where the water table is below the land surface. A subsurface map of pore water velocities (V), which is required to perform subsurface water particle tracking, was obtained as:

$$V_m^x\,(x,y,z) = \frac{1}{\theta_s} q_m^x\,(x,y,z) \tag{S.4a}$$

$$V_m^y\,(x,y,z) = \frac{1}{\theta_s} q_m^y\,(x,y,z) \tag{S.4b}$$

$$V_m^z\,(x,y,z) = \frac{1}{\theta_s} q_m^z\,(x,y,z) \tag{S.4c}$$

where θ_s is subsurface porosity. Inputs to the model included: (1) the location and water level of wetlands; and (2) material properties of the subsurface. For (1), the delineated wetlands explained in Sect. 2.2 were used, and we assumed that the water level was equal to the average elevation of each wetland boundary. This water level was used as a constant head boundary condition at each wetland. For (2), a two-layer unconfined aquifer with a 5 m thick shallow layer was used to characterize the subsurface (as suggested in van der Kamp and Hayashi, 2009). The bottom boundary of the computational domain was assumed to be at Z = 0 with a no-flow condition. No-flow side boundaries were also placed on average 20 km away from the watershed border; this treatment minimized the impact of side boundaries on flow behaviour and subsurface connections. A porosity (θ_s) value of 0.14 equal to the average measured porosity at the Beaverhill watershed (Gleeson et al., 2014) was also used.