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Integration of models using shared state variables: Implementation in the regional hydrologic modelling system SIMGRO

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SUMMARY

SIMGRO is an integration hub that connects to diverse hydrologic models. Its coupling method aims to describe the hydrologic feedbacks adequately at an acceptable cost of the increased computational load. The achieved balance between accuracy and efficiency is investigated in the paper. The coupling method centres around a shared state variable of two connecting subsystems. A prime example of such a variable is the elevation of the phreatic surface that forms the natural demarcation between the saturated groundwater flow and the unsaturated flow in the soil. In our scheme the shared variable is alternately updated by the connecting submodels, with both models using the same combined storage relationship. This method has also been implemented for the linkage between the ponding water of column models and surface water models. Using a basin-scale model, the scheme was subjected to a time-step sensitivity analysis. Results were compared to a benchmark, for which we took the same scheme, but with a short time step of 0.125 d. Increasing the time step to 1 d caused a numerical error of less than 5% in the simulated regional design discharge. Errors in the simulated 95th percentile of highest groundwater levels are less than 0.05 m. The coupling method presented here is a key element of the Netherlands Hydrological Modelling Instrument. The current version involves half a million spatial units of 250 × 250 m.

1. Introduction

Integrated hydrologic models are useful tools for simulating effects of water management measures in situations where feedbacks play a role. Their development started in the late 1960s. The used conceptualizations are very diverse. Moreover, the used coupling methods differ widely in the way submodels are linked up and in the way the time stepping is organized.

The weakest form of model integration is provided by the 'sequential' coupling scheme: first Model A performs a time step with a Dirichlet boundary condition from Model B; the calculated flux is then inserted as a Neumann boundary condition of Model B, and the latter is updated for the time step. An example of this approach would be the coupling between Hydrus-1D and MODFLOW (Twarakavi et al., 2008; Seo et al., 2007): Hydrus-1D first uses the MODFLOW head at the bottom of the soil profile as a boundary condition; subsequently MODFLOW uses the flux simulated by Hydrus-1D as a recharge rate. The disadvantage of such a sequential coupling scheme is that the flux between the two model domains is completely determined by the model that gets updated first, which can lead to instability of the model that comes after it. Remedying this instability of a sequential scheme via reduction of the time step provides a poor balance between accuracy and efficiency (Clark and Kavetski, 2010).

The ultimate form of integration is provided by embedding the submodels in a single set of equations that can be solved with an off-the-shelf mathematical package. This 'fully coupled' scheme guarantees full feedback (between submodels) within each time step. An example of such an approach would be the model for surface and subsurface flow HydroGeoSphere (Therrien et al., 2010). The main drawback of a fully coupled model is that combining the equations of the submodels into a single set increases its computational load exponentially. This is also known as the 'curse of dimensionality'. If that leads to inefficiency, an alternative is to iteratively solve the equations of a sequential scheme, with the flux computed by Model A using the updated Dirichlet boundary conditions from Model B. This method is called *iterative coupling* by Furman (2008, Fig. 2). Examples of such a type of coupling to MODFLOW are the unsaturated zone package VSF of Thoms et al. (2006) and the stream package of Prudic (1989).

Iterative coupling can put a large demand on computational resources due to the computational expense of an iteration (VSF) or because a high conductance between submodels hinders convergence. Such convergence problems are not experienced with the 'shared-variable' scheme of SIMGRO (Van Walsum et al., 2011). The method uses a shared state variable for the link-up between SVAT models

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(Soil Vegetation Atmosphere Transfer) and groundwater and surface water models. It has been implemented in the Netherlands Hydrologic Modelling Instrument (Delsman et al., 2008). The first version of this instrument has a resolution of over half a million grid cells. Regional implementations are also underway, for instance the model for the waterboard 'De Stichtse Rijnlanden'.

In the paper, a short overview of SIMGRO is given in the methodology section. The second part of the section describes in more detail how the coupling method has been implemented. In the results section, the sensitivity of the simulations for the used time step is investigated and evaluated. In the conclusions the achieved performance of the method is summarized.

2. Model coupling method and implementation

2.1. Overview of SIMGRO

The SIMGRO modelling framework (Querner, 1988; Veldhuizen et al., 1998; Van Walsum et al., 2011) connects to:

- a SVAT column model for the soil–water–vegetation–atmosphere interactions, for which MetaSWAP is used (Van Walsum and Groenendijk, 2008);
- a groundwater model, for which MODFLOW is used (McDonald and Harbaugh, 1988);
- a surface water model.

Model interactions take place within two nested time cycles:

- the cycle for 'fast' hydrologic processes: interactions with the atmosphere, the soil surface processes including infiltration, and the interactions between the SVAT models and the surface water model(s);
- the cycle for the 'slow' hydrologic processes: unsaturated flow in the SVAT models and the groundwater flow.

Some form of geo-referencing is required for codifying links between model units. For this, the European standard Open Modeling Interface – 'OpenMI' (HarmonIT, 2005) – has various formats. In OpenMI terminology, the SIMGRO connections between the submodels are '*id*-based', meaning that the connections are simply defined in terms of mapping tables containing links between identifiers of model units. These links do not have to be on a one-to-one basis: there can be multiple units of one model coupled to a single unit of the other model.

Coupling methods can differ in the way schematizations of submodels are linked up and in the way the time stepping is organized. Most schematizations involve a discontinuity of the system state variable across the submodel boundaries. For continuous state-space schematizations, the link-up between two submodels can also be done via a shared variable describing the system state on the subsystem boundary. This method has been implemented in SIMGRO for the coupling between SVAT models and a groundwater model, and for the coupling between the ponding water of SVAT models and a surface water model.

2.2. Integrated modelling of groundwater and soil moisture

The simplest way to link SVAT models and a groundwater model is through a flux link across a fixed boundary, like is done in the coupling between Hydrus-1D and MODFLOW (Twarakavi et al., 2008; Seo et al., 2007). Such a flux-link coupling is indicated by the '*q*-link' on the right-hand side of Fig. 1, with the model boundary taken below the lowest groundwater level that can occur. In the SIMGRO implementation of this method, the flux is computed via a

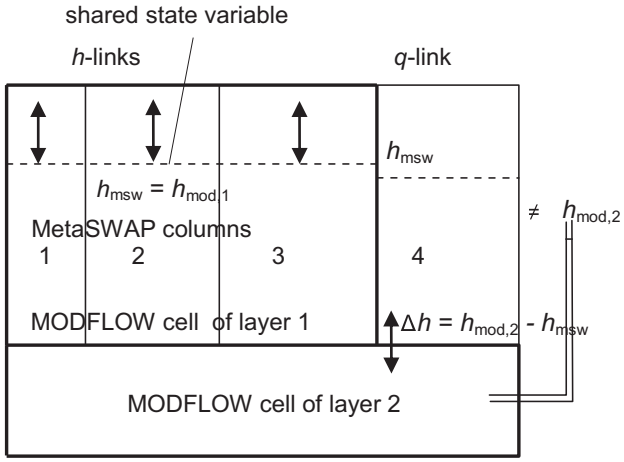

Fig. 1. Cross-section of the MetaSWAP-MODFLOW coupled systems, showing examples of so-called *h*-links via a shared state variable (phreatic level h_{msw}) on the left, and a flux link ('*q*-link') on the right.

'resistance' relationship: $q = \Delta h/c_{bot}$, where q is the interaction flux ($m\ d^{-1}$), Δh is the head difference between the phreatic surface of the SVAT model and the connecting MODFLOW cell (m), and c_{bot} is the flow resistance (d) (which is the inverse of the MODFLOW-style "conductance" per unit of area). The flux is iteratively updated in the "outer loop" of the MODFLOW solution cycle.

The MODFLOW-SVAT coupling method with a flux link has the drawback that the MODFLOW model is 'buried' below the SVAT models. The consequence is that the horizontal flow in the saturated part of the SVAT columns can not be modelled with the MODFLOW functionality.

This schematization problem of the flux-link method can be avoided by locating the model interface just below the phreatic level, and then letting it follow like a shadow. In the 'limit' of this schematization, the distance of the boundary below the phreatic level becomes infinitesimally small, and the resistance c of the flux-relationship approaches zero. However, the low resistance makes it hard to get the iteration cycle to converge, thus making the method very inefficient.

As a numerical equivalent of the above described moving-boundary schematization, a scheme was developed that uses the phreatic level itself for the link-up between the SVAT model and MODFLOW. In this method, the phreatic level is a 'shared state variable' that is present in the model equations of both the SVAT model and the groundwater model. In short, this is called the *h*-link method. Such links are present in the left part of Fig. 1, where an example is given of the *N*:1 linkage of 3 SVATs to a MODFLOW cell. The 3 SVATs all have the same phreatic level, because they all share this state variable with the head of a single MODFLOW cell.

In high-resolution regional applications, the Richards-type model puts an excessive demand on computational resources. That was the reason for developing MetaSWAP (Van Walsum and Groenendijk, 2008). It is based on a quasi-steady state schematization of the flow processes in combination with water balances for control volumes at aggregate scale. The balances include a storage change term; this makes the model dynamic. Application of the model involves three phases:

- pre-processing of basic soil physical data, including an aggregation step;
- simulation of multiple MetaSWAP models coupled to MODFLOW;
- post-processing of the MetaSWAP model results, including a disaggregation step that produces moisture profiles for each time step (optional).

In the pre-processing phase with the steady-state version of the SWAP model (Van Dam et al., 2008), a series of simulations is done for a set of phreatic levels and atmospheric water demands. A large number of steady-state moisture profiles are produced. For each of them the mean root zone pressure head ψ_r is derived. The ψ_r-value in combination with the groundwater level h is unique for each of the profiles. That makes it possible to use ψ_r and h as the independent variables of 'meta'-functions describing the storage and flow characteristics of the profiles:

- **tbs**$_l(\psi_r, h)$, for the storage in an aggregation layer l, as a function of the groundwater level, h, and the mean pressure head in the root zone, ψ_r;
- **tbq**(ψ_r, h), for the unsaturated flux.

An example of a root zone storage metafunction is given in Fig. 2.

The simulation method is here explained for a model implementation with just one aggregation layer. For more than one layer (three is considered minimum) the details can be found in Van Walsum and Groenendijk (2008). The simulation method is first explained for a stand-alone column with no external stresses on the saturated part of it; this is followed by the explanation for the coupled MetaSWAP-MODFLOW combination.

In a stand-alone model application of MetaSWAP without any externally imposed saturated fluxes, a time step consists of two substeps:

- update of the root zone pressure head, assuming an unchanged phreatic level;
- update of the phreatic level.

In the first substep, the update of the root zone pressure head is done in such a manner that the recharge entering the profile (infiltration minus soil evaporation and transpiration) is balanced by the storage change plus the unsaturated flux to the phreatic level.

In the preparation for the second substep, the new root zone pressure head is then inserted into the bivariate storage functions **tbs**$_l(\psi_r, h)$; summation over the layers yields the monovariate function **tbs**$^{\sim}(h)$. It gives the relationship between the phreatic level and the total storage in the control volume of the vertical profile. An example of such a function is given in Fig. 3a.

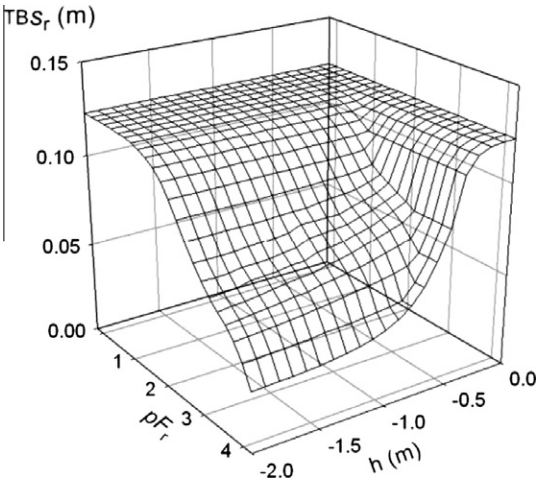

Fig. 2. Example of a bivariate tabular root zone storage function **tbs**$_r(\psi_r, h)$ for a loamy soil with a root zone thickness of 0.30 m. The total storage in the root zone is a function of the mean pressure head in the root zone (pF of ψ_r) and the phreatic level (h).

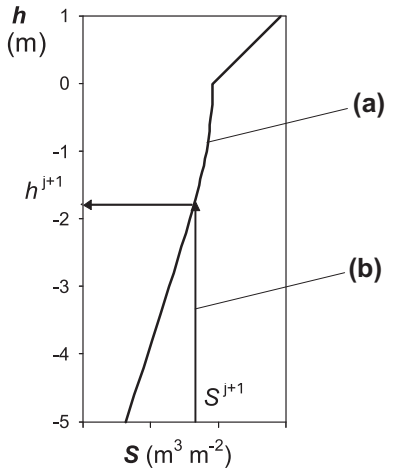

Fig. 3. (a) Example of a monovariate storage relationship, giving the total storage S in the vertical profile as a function of the phreatic level h. The relationship is derived by inserting the prevailing root zone pressure head in the bivariate storage functions as shown in Fig. 2, and then summating over all the aggregation layers of the vertical profile. The datum plane of h is at the soil surface; above the soil surface the storage of the ponding water is equal to the head. (b) Use of the relationship for the update of the phreatic level, h^{j+1}, based on the newly computed total storage of water in the profile, S^{j+1}.

In the second substep (with no external saturated fluxes imposed in this case), the following water balance equation is evaluated for the vertical profile:

$$S^{j+1} = S^{j} + R\Delta t \tag{1}$$

where S^{j}, S^{j+1} is the storage in the vertical profile at time level j and $j+1$, respectively (m^3 m^{-2}), R the time-averaged recharge of time interval (m^3 m^{-2} d^{-1}), Δt is the time step of soil moisture/groundwater models (d).

The evaluated S^{j+1} can then be used for an inverse look-up operation in the tabular storage function **tbs**$^{\sim}(h)$, as shown in Fig. 3b. This yields the update of the phreatic level by the MetaSWAP model, h^{j+1}.

In the numerical scheme for the coupling of MetaSWAP and MODFLOW, the phreatic level is modelled as a shared state variable that is influenced by both the unsaturated and the saturated flow. These influences are implemented by letting both models update the phreatic level, one after the other.

The update of the phreatic level by MetaSWAP is communicated to MODFLOW in terms of a *flux*. That flux is not the same as the recharge given in Eq. (1), because it must take into account that part of R has been used by MetaSWAP for the update of the root zone pressure head and the associated storage change. The part of the recharge that is left over is available for the phreatic level update and the associated storage change:

$$q_{msw} = R - (S^{\sim j+1} - S^{j})/\Delta t \tag{2}$$

where q_{msw} is the time-averaged unsaturated flux (m^3 m^{-2} d^{-1}) and $S^{\sim j+1}$ is the storage of the profile after the first substep of the MetaSWAP model (m^3 m^{-2}), equivalent to **tbs**$^{\sim}(h^{j})$. The water balance update of the complete vertical profile can now be written as:

$$S^{j+1} = S^{\sim j+1} + (q_{msw} + q_{mod})\Delta t \tag{3}$$

where q_{mod} is the nett MODFLOW flux (m^3 m^{-2} d^{-1}), the unknown that is still to be computed by the MODFLOW model. In the case that $q_{mod} = 0$, the new S^{j+1} reverts to the value given by Eq. (1), as can be seen by inserting the expression for q_{msw} given in Eq. (2) into Eq. (3). The total change of the phreatic level is then only determined by the MetaSWAP update of it.

The balance given by Eq. (3) is for a *control volume* that comprises both the saturated and unsaturated zone. This way of conserving mass should not be confused with a balance for groundwater as a *system volume*. If the system volume approach had been used here, then any change of the groundwater level would strongly influence the exchange flux between the unsaturated and saturated part of the profile, because for instance a rising groundwater level 'eats up' water in the unsaturated zone as it rises. Then it would be erroneous to determine q_{msw} in a 'sequential' substep as is done here. It should then be part of an iterative coupling scheme.

Apart from the unsaturated flux, MetaSWAP also communicates to MODFLOW the associated storage response for changes in the MODFLOW head, under the assumption that the MetaSWAP phreatic level is forced to follow the update of the MODFLOW head. This is done in the form of the tabular storage relationship $\mathbf{tbs}^{\sim}(h)$. The relationship includes a prediction of how the root zone pressure head itself will be influenced by any change of the phreatic level that is imposed by an externally coupled model; this makes the scheme stable and more accurate.

MODFLOW combines the processing of the MetaSWAP information with its update of the linked MODFLOW head. In the numerical implementation, a complication is caused by the fact that MODFLOW does not have its own package for handling a nonlinear storage relationship. Therefore, MetaSWAP supplies the MODFLOW model with a storage coefficient of the phreatic layer, for each 'outer loop' cycle of the PCG-solver. For deriving new values of the coefficients, use is made of the latest information from MODFLOW about the fluxes and heads. Depending on the situation, the update is either done as the 'level-based' or the 'balance-based' storage coefficient, as defined in Van Walsum and Groenendijk (2008, Fig. 10). These storage coefficients serve the purpose of telling MODFLOW how the total storage will be affected by changing the shared state variable, i.e. the MODFLOW head that is linked 1:1 to the MetaSWAP groundwater level. In MODFLOW, Eq. (3) is thus implemented in the form of:

$$\mu(h_{mod}^{j+1} - h_{mod}^{j}) = (q_{msw} + q_{mod})\Delta t \tag{4}$$

where μ is the storage coefficient derived from the storage table $\mathbf{tbs}^{\sim}(h^j)$ ($m^3\,m^{-2}\,m^{-1}$), h_{mod}^{j} is the MODFLOW head at time level j (m). At all times the water balance integrity of the coupled models is guaranteed by Eq. (3), which is administrated in the MetaSWAP model. The storage change computed by MODFLOW with Eq. (4) only plays a role in letting the MODFLOW head and MetaSWAP groundwater level converge, so that MODFLOW simulates a q_{mod} flux for a head that is (nearly) equal to the MetaSWAP groundwater level. The final value of the MODFLOW flux is used for finalizing the soil moisture profile of MetaSWAP. An overview of the solution procedure is given in Fig. 4.

A convergence criterion is used for stopping the iteration process, in combination with a maximum to the number of iterations. In a big model with the maximum number of iterations set at 5, there typically are residual head differences of $\geqslant$0.01 m in 0.1% of the SVAT columns. Any residual head difference is interpreted as a deviation from the zero flow resistance that is implied by the 1:1 coupling of the MODFLOW head and the MetaSWAP groundwater level. A method has been implemented for avoiding a subsequent water balance error during the next time step (Van Walsum et al., 2011).

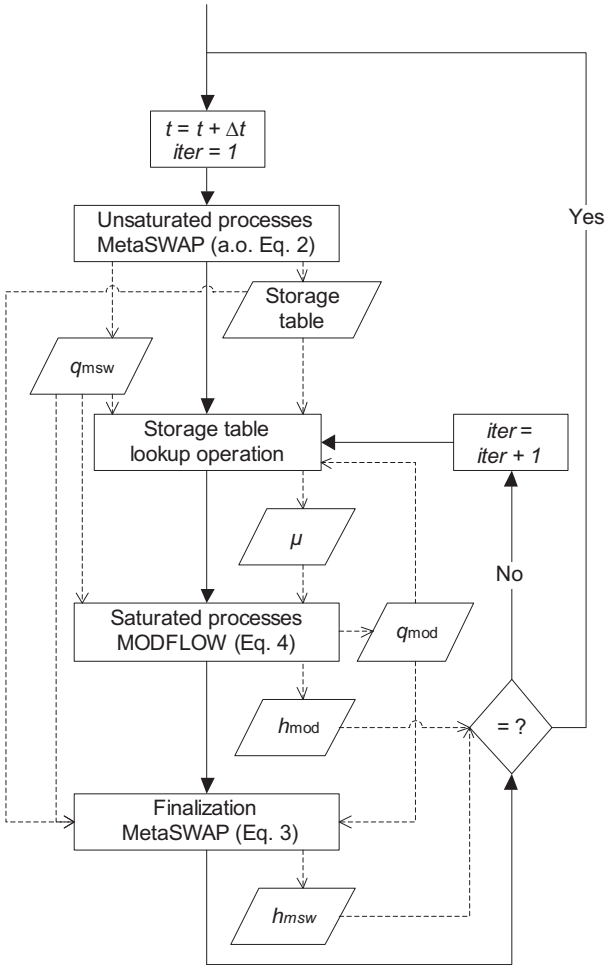

Fig. 4. Overview of the solution cycle for the coupled MetaSWAP-MODFLOW models. Explanation of used symbols: q_{msw} is the unsaturated flux ($m^3\,m^{-2}\,d^{-1}$), q_{mod} is the saturated flux ($m^3\,m^{-2}\,d^{-1}$), μ is the storage coefficient "sc1" of MODFLOW's top layer ($m^3\,m^{-2}\,m^{-1}$), h_{mod} is the MODFLOW head of the top layer (m), and h_{msw} is the groundwater level in MetaSWAP (m). Continuous lines show the flow of computational control; dashed lines show the flow of data.

2.3. Integrated modelling of ponding water and channel surface water

For over-the-soil-surface interactions between ponding water and surface water, a flux-link scheme is computationally very demanding, due to the low resistance of the flux-link relationship. To keep it stable will require a very short time step. An alternative is then to use a shared state variable for the coupling.

The *h*-link coupling of ponding water to a surface water model can be implemented if the lateral head differences between the ponding water and the channel surface water are assumed negligible. An example of such a situation is shown in the floodplain cross-section given in Fig. 5. For the given assumption, the cross section has a water level that is specified by a single state variable. In the example there is only one SVAT column. But in real applications there will be several SVAT columns linked to a section of a surface water model.

The coupling scheme follows the same general pattern as the flow chart given for the MetaSWAP-MODFLOW coupling (Fig. 4), but without the iteration loop for the updates of the storage coefficient. In the coupling to the surface water model, the storage tables of the ponding water can be communicated to the surface water model before the simulation actually starts, because the tables themselves do not change: they are purely determined by the soil surface topography. The surface water model uses the combined storage tables with the nonlinear storage relationships within its implicit solution scheme. The iteration loop for the storage coefficient is not needed here, because a surface water model can cope with a nonlinear storage relationship, which MODFLOW is not able to do.

Running the coupled models involves the following steps (Fig. 5):

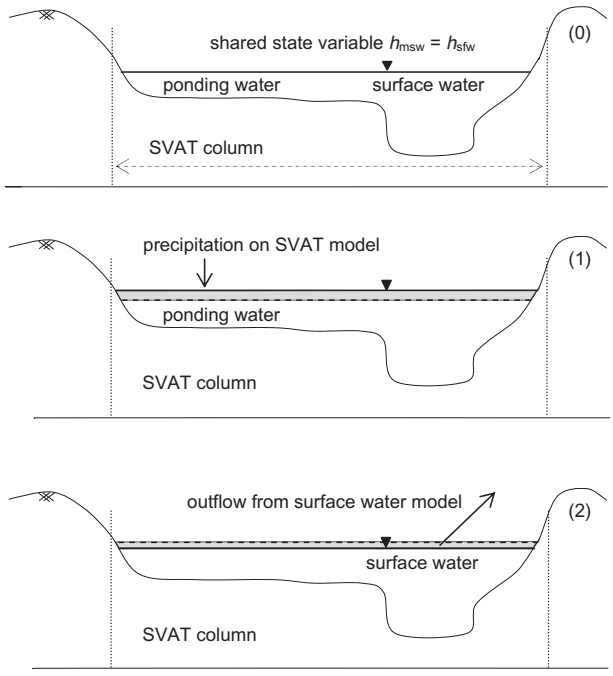

Fig. 5. Computational steps of integrated model for ponding water (h_{msw}) and surface water (h_{sfw}).

0. Initial state, with ponding/inundation of the flood plain.
1. The SVAT columns simulate the nett recharge of the ponding water; these updates are communicated to the surface water model.
2. The surface water model simulates outflow, which in this case leads to a lowering of the level; changes are communicated to the SVAT models.
3. The levels in the SVAT models are updated for the changes made by the surface water model; the nett fluxes from the SVAT columns to the surface water model are determined from the water balances of the SVAT columns.

2.4. Evaluation method for numerical performance of time stepping scheme

In the Netherlands, a method has been developed to estimate the crop damage from the so-called 'Mean Highest' water table, *MHW*. The latter roughly coincides with the 95th percentile of the cumulative probability distribution of daily groundwater levels, h_{95}, with the datum plane taken at soil surface. For the evaluation methods based on the 95th percentile, the numerical error in the groundwater level simulation should not be more than 0.05 m. As a benchmark we used the results for a time step of 0.125 d.

In the Netherlands, the peak flow (averaged over a time span of 2 d) with a return period of 1 year is commonly used for designing channel networks. For evaluating its accurate simulation, the used criterion is that the simulated discharge for a region as a whole should not differ more than 5% from the results for the benchmark.

3. Results and discussion

3.1. Coupling between MetaSWAP and MODFLOW

3.1.1. Verification of implementation

The validation of the unsaturated zone concept of MetaSWAP was reported in Van Walsum and Groenendijk (2008). Here the *h*-link coupling between MetaSWAP and MODFLOW is first investigated for its numerical correctness, to provide a verification of its implementation. This verification of the *h*-link coupling is performed by comparing results to a supposedly equivalent formulation in terms of *q*-links.

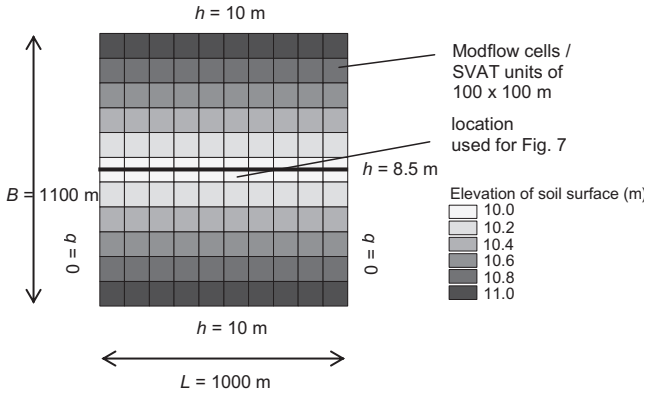

Fig. 6. Model schematization for testing the *h*-link coupling between MetaSWAP and MODFLOW. The MODFLOW model has been implemented with the following packages and following parameters: (i) block-centred package (bcf) with conductance of 5 $m^2\,d^{-1}$, head boundary of 10 m, (ii) river package with conductance of 400 $m^2\,d^{-1}$, water level of 8.5 m. The MetaSWAP model has been implemented with the following parameters: (i) tree vegetation with root zone depth of 1.0 m, (ii) loamy soil.

To do this, a simple MODFLOW-MetaSWAP model was set up for a rectangular region as depicted in Fig. 6. Two versions of the model were implemented: one with *h*-links, the other with *q*-links.

The *q*-link version uses a flux relationship of the following form:

$$q_{mod} = (h_{mod} - h_{msw})/c_{bot} \quad (5)$$

where: q_{mod} is the flux from the MODFLOW model to the MetaSWAP model (m d^{-1}); h_{mod} is the MODFLOW head of the top layer (which is the only layer in the example model) (m); h_{msw} is the phreatic level in the MetaSWAP model (m); c_{bot} is the 'bottom flux' resistance (d).

For the implementation with the *q*-links to become equivalent to the one with the *h*-links, the bottom flux resistance should approach zero; here that is approximated by using a value of 0.05 d.

In the *q*-link option, the top layer of the MODFLOW model becomes 'confined'; for the results to approach the results with the *h*-link there should be no additional storage in the MODFLOW model; this has been approximated by setting the storage coefficient to 10^{-5} ($m^3\,m^{-2}\,m^{-1}$).

For each outer loop cycle of MODFLOW, the interaction flux q_{mod} is updated. Keeping this iteration process stable required special measures:

- an under-relaxation factor of 0.0005 during the first 500 outer loops, increasing to 0.003 for the outer loops from 500 to 1000, and then above 1000 increasing to 0.01;
- a time step of 1 h.

In Fig. 7 a comparison is given between the simulated phreatic levels of MetaSWAP for the two types of model coupling. It shows that the simulation with a *q*-link lags slightly behind that of the *h*-link, and that it also has a slightly smaller total fluctuation. This is attributed to the combined effect of the non-zero c_{bot} resistance and the non-zero added storage of the MODFLOW model. In terms of the RMSE (Root Mean Squared Error) the difference is 0.022 m; the maximum difference is 0.087 m, with the *q*-link coupling producing a lower maximum level than the *h*-link one. If a lower resistance c_{bot} were to be used in combination with a smaller time steps (and an appropriate under-relaxation scheme), we consider it safe to assume that results for the *q*- and *h*-link coupling would con-

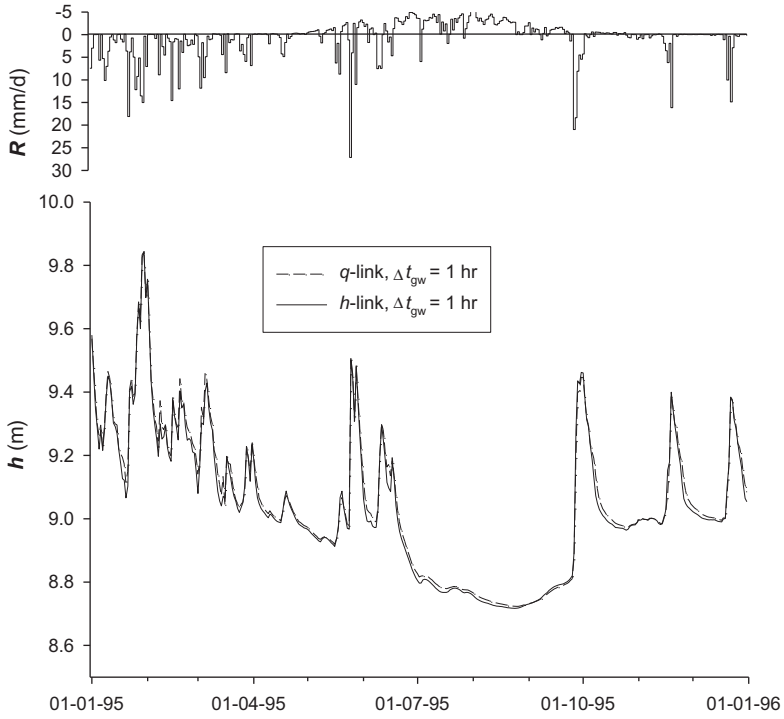

Fig. 7. Comparison between simulation results for the phreatic level h, using the *q*-link and the *h*-link coupling method, for the location in the test model indicated in Fig. 6. The plotting time interval is 1 d. The recharge *R* has been taken as the infiltration at the soil surface minus the actual transpiration.

Fig. 8. Simulated map of the 95th percentile of groundwater levels (with the soil surface as datum plane; the cumulative distribution was derived from down-negative values), for the area of "Waterboard Stichtse Rijnlanden" (www.hdsr.nl). The nett model area (shown) is 82,000 ha. The gross model area is more than twice the size (190,000 ha), for the purpose of keeping the model boundary condition at a 'safe' distance so that it does influence the area that is of interest. For the SVAT-modelling 325,000 MetaSWAP columns have been coupled to a MODFLOW model that has a 100 × 100 m grid and eight model layers.

verge to the same values, thus implying that the *h*-link coupling has been implemented in a correct manner.

3.1.2. Numerical accuracy of time stepping scheme

A sequential coupling scheme has a so-called 'splitting error' (Clark and Kavetski, 2010; Schoups et al., 2010). The numerical accuracy of the time stepping scheme was investigated by performing a sensitivity analysis for the used time step. This was a done with the model of waterboard 'De Stichtse Rijnlanden' (<www.hdsr.nl>). The model has been parameterized in such a manner that the surface runoff simulation is done by the drainage package of MODFLOW, with the drainage base (of the surface runoff records) at the elevation of the soil surface.

The accuracy of the shared state variable scheme is investigated for simulating the 95th percentile of the groundwater level, and for predicting the peak flow Q_{des} with a 1 year return period (Section 2.4). The simulations were performed for the period 1989–2006, thus comprising 16 complete 'hydrologic' years. This period was simulated with time steps of respectively 0.125 d (3 h), 0.25 d, 0.5 d, and 1 d.

Table 1
Comparison between simulated maps of the 95th percentile of the cumulative frequency distribution *F* of the groundwater level *h*, using time steps of 0.125 d, 0.25 d, 0.5 d and 1 d. The map simulated with a time step of 0.125 d is taken as a reference. The used groundwater levels are at the end of a day.

Δt (d)	$\lvert\Delta h\rvert \leqslant$ 0.015 m (area %)	0.015 m $\leqslant \lvert\Delta h\rvert <$ 0.035 m (%)	0.035 m $\leqslant \lvert\Delta h\rvert <$ 0.055 m (%)	0.055 m $\leqslant \lvert\Delta h\rvert$ (%)
0.25	98.7	1.3	0.0	0.0
0.50	92.3	7.4	0.1	0.2
1.00	76.9	21.6	1.1	0.3

Table 2
Comparison between simulated maps of the design discharge Q_{des}, with an average return period of 1 year. The results for a time step of 0.125 d is taken as the reference. For this the simulated design discharge of the region as a whole is Q_{des} = 7.90 mm d^{-1}. For larger time steps the deviations are given in the column $\Delta Q_{des}/Q_{des}$. The deviations given in the column $\Delta Q_{max}/Q_{max}$ are for the maximum flow in the 16 year simulation period; for a time step of 0.125 d the simulated value is Q_{max} = 11.4 mm d^{-1}.

Δt (d)	$\lvert\Delta Q\rvert \leqslant 1$ mm d^{-1} (area %)	$1 \leqslant \lvert\Delta Q\rvert < 2$ mm d^{-1} (area %)	$2 \leqslant \lvert\Delta Q\rvert < 5$ mm d^{-1} (area %)	$\Delta Q \geqslant 5$ mm d^{-1} (area %)	$\Delta Q_{des}/Q_{des}$ (%)	$\Delta Q_{max}/Q_{max}$ (%)
0.25	>99.9	<0.1	–	–	1.0	<0.1
0.50	98.5	1.4	<0.1	–	2.4	−0.1
1.00	85.1	14.0	0.8	<0.1	3.5	−1.8

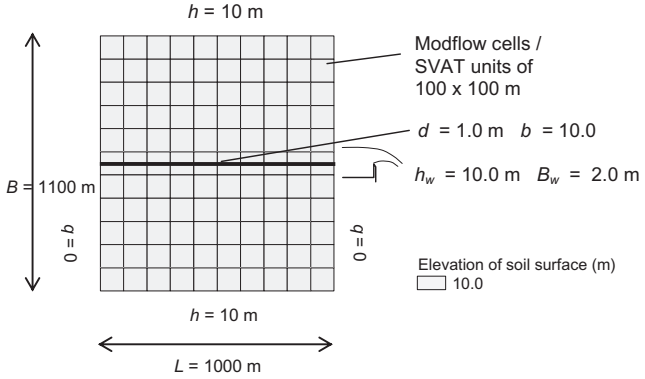

Fig. 9. Model schematization for testing (numerical implementation) of *h*-link coupling between MetaSWAP and a surface water model. The MODFLOW model has been implemented with the block-centred package (bcf) with conductance of 5 m^2 d^{-1}, head boundary of 10 m. The MetaSWAP model has been implemented with a tree vegetation with root zone depth of 1.0 on a loamy soil. The SIMGRO surface water module has been implemented with: (i) a channel with depth (d) of 1 m, breadth (b) of 10 m, and a groundwater-surface water conductance of 400 m^2 d^{-1}, (ii) a weir with a crest level (h_w) at 10 m and a breadth (B_w) of 2 m.

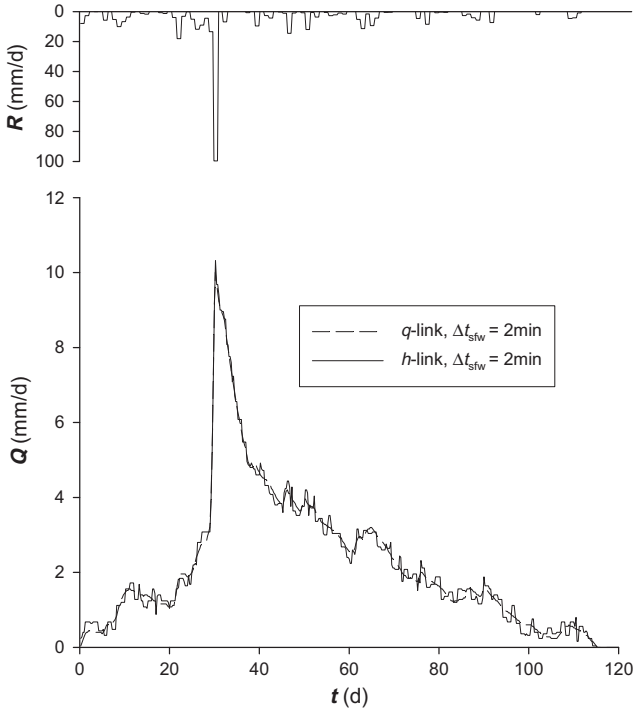

Fig. 10. Comparison between simulated peak flows (time averaged over 0.25 d) using the model of Fig. 9 with respectively an *h*-link and a *q*-link with a runoff resistance *c* of 0.083 d. The used surface water time step is 2 min. The recharge *R* of the model is here defined in terms of the total precipitation minus the total actual evapotranspiration.

The h_{95}-map simulated with a time step of 0.125 d (Fig. 8) was taken as the reference for making cross comparisons with results for other time steps. The comparison to the maps simulated for larger time steps is summarized in Table 1. From these results it appears that a time step of 1 d does not cause a significant area percentage of the model to have a numerical error greater than the 0.05 m that is considered necessary for accurate application of the evaluation methods (Section 2.4).

Like for the h_{95} map, the model run with a time step of 0.125 d was taken as the reference for the simulation of the peak flow Q_{des} with a 1 year return period. For obtaining the Q_{des} values, the results for the 0.125 d time steps were averaged over 2-d intervals. The comparison to the maps simulated for larger time steps is summarized in Table 2. The design discharge increases slightly with increasing time step. That is due to the splitting error introduced by simulating the unsaturated flux in a sequential step that is not updated in the implicit part of the scheme (Fig. 4). For the larger time step the feedback from the rising groundwater level to the simulated percolation is less than the 'full feedback' that is simulated with a time step of 0.125 d. But for the total discharge of the region as a whole this is not a significant numerical error. The influence of the time step on the maximum discharge (last column of Table 2) is even less. The peak includes 8% of saturation-induced runoff.

3.2. Coupling between ponding water in MetaSWAP and a surface water model

3.2.1. Verification of implementation

The numerical implementation of the *h*-link coupling of MetaSWAP to a surface water model was tested with a synthetic model of a flood plain situation (Fig. 9). The flood plain dynamics were simulated for a meteorological time series that includes a precipitation event of 100 mm within a single day. This event was simulated using the following time stepping:

- Δt_{gw} = 0.25 d for MODFLOW and MetaSWAP;
- Δt_{sfw} = 2 min for the surface water model and for its interactions with the ponding water of MetaSWAP.

The model was implemented in two versions, using respectively *h*-links and *q*-links between MetaSWAP and the surface water model. The *q*-links make use of the following relationship:

$$q_{run} = (h_{sfw} - h_{pd})/c_{run} \quad (6)$$

where: q_{run} is the runoff flux from the MetaSWAP model to the surface water model (m d^{-1}); h_{sfw} is the surface water level (m); h_{pd} is the ponding level in the MetaSWAP model (m); c_{run} is the runoff resistance (d).

The *q*-links were implemented with respectively a resistance of 0.25 d, 0.125 d, and 0.083 d (2 h). The simulated peak flow over a 0.25 d time interval (time-averaged value) was found to be respectively 9.88, 9.96 and 10.00 mm d^{-1}. With the *h*-link version of the model, a peak flow of 10.32 mm d^{-1} was simulated. In Fig. 10 the comparison is made to the simulation for a *q*-link with a resistance of 0.083 d. With the latter taken as the benchmark, the RMSE of

the *h*-link simulation is 0.18 mm d^{-1}. Such a low value justifies the conclusion that the *h*-link coupling has been correctly implemented.

3.2.2. Numerical accuracy of time stepping scheme

The sensitivity of the *h*-link coupling for the used surface water time step was investigated by comparing results for a time step of 2 min and 1 h. The simulated peak flows were respectively 10.32 and 10.12 mm d^{-1}, for a time-averaging interval of 0.25 d. For this time-averaging interval, situations with fast-reacting runoff can apparently be accurately simulated with an *h*-link using time step of 1 h. In regional model applications, a time step of 1 h for the 'fast' processes is often a good compromise between the desire to confine the error caused by a too large time step and also to keep the total simulation time within practical limits.

4. Conclusions

Model coupling with existing methods is problematic if the flow resistance across the boundaries of subdomains is small (i.e. the conductance is very high), like in the case of runoff from an inundated flood plain to a river. In the case of saturated–unsaturated flow, the phreatic level forms the natural demarcation of the subdomains, and any resistance that is introduced between them is a model artefact.

The method presented avoids the use of a resistance between submodels, and replaces it by introducing the concept of a shared state variable. This shared state variable is the pivotal element in a 'sequential' coupling scheme. In the terminology of Furman (2008, Fig. 2), the scheme is 'uncoupled'. Normally such a scheme has a poor balance between accuracy and efficiency. The novelty of our scheme is that the exchange of information between the models not only includes a flux, but also a (nonlinear) storage relationship. This storage relationship can be seen as a boundary condition of the 'third' kind.

In the MetaSWAP to MODFLOW coupling, the scheme has been implemented in the form of iterative storage-coefficient updates that are fully integrated in the implicit MODFLOW scheme. In the case of the MetaSWAP to surface-water-model coupling the coupling has been done by exporting the storage relationships of ponding water to the surface water model in the initialization phase of the simulation. By varying the used modelling time step and comparing the results to those for a short time step it was shown that the scheme has an accuracy that nearly equals that of an 'iteratively coupled' scheme for the mentioned model couplings.

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