



## Assessing the influence of groundwater and land surface scheme in the modelling of land surface-atmosphere feedbacks over the FIFE area in Kansas, USA

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# Environmental Earth Sciences

## Assessing the influence of groundwater and land surface scheme in the modelling of land surface-atmosphere feedbacks over the FIFE area in Kansas, USA --Manuscript Draft--

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<b>Order of Authors Secondary Information:</b>	
<b>Abstract:</b>	<p>The land surface-atmosphere interaction is described differently in large scale surface schemes of regional climate models and small scale spatially distributed hydrological models. In particular, the hydrological models include the influence of shallow groundwater on evapotranspiration during dry periods where soils are depleted and groundwater is the only water supply. These mechanisms are analysed by combining a distributed hydrological model (MIKE SHE) and a regional climate model (HIRHAM) and comparing simulation results to the FIFE area observation data in Kansas, USA. The numerical experiments include five simulations. First MIKE SHE is forced by observed climate data in two versions i) with groundwater at a fixed uniform depth, and ii) with a dynamical groundwater component simulating shallow groundwater conditions in river valleys. iii) In a third simulation MIKE SHE is forced by HIRHAM simulated precipitation. The last two simulations include iv) a standard HIRHAM simulation, and v) a fully coupled HIRHAM-MIKE SHE simulation locally replacing the land surface scheme by MIKE SHE for the FIFE area, while HIRHAM in standard configuration is used for the remaining model area.</p> <p>The results show a clear correlation between depth to the groundwater and evapotranspiration with a distinct groundwater depth threshold at 0.5-3 m. During the dry summer period the two MIKE SHE simulations using distributed groundwater reproduced evapotranspiration better than MIKE SHE with unsaturated flow alone and the HIRHAM simulations. This indicates that including dynamic groundwater in a fully coupled climate-hydrological model may improve evapotranspiration fluxes from areas with shallow groundwater tables.</p>
<b>Suggested Reviewers:</b>	Jehan Rihani, Ph.D. Scientist, University of Bonn jrihani@uni-bonn.de Knowledge on land surface/atmosphere interaction also in North American settings

	Reed Maxwell, Ph.D. Professor, Colorado School of Mines maxwell@mines.edu High degree of expertise in land surface/groundwater/atmosphere interaction.
<b>Opposed Reviewers:</b>	

# 1 **Assessing the influence of groundwater and land surface scheme in the** 2 **modelling of land surface-atmosphere feedbacks over the FIFE area in** 3 4 **Kansas, USA** 5

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## 23 **Abstract**

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26 The land surface-atmosphere interaction is described differently in large scale surface schemes of regional  
27 climate models and small scale spatially distributed hydrological models. In particular, the hydrological  
28 models include the influence of shallow groundwater on evapotranspiration during dry periods where soils  
29 are depleted and groundwater is the only water supply. These mechanisms are analysed by combining a  
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33 fixed uniform depth, and ii) with a dynamical groundwater component simulating shallow groundwater  
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36 HIRHAM-MIKE SHE simulation locally replacing the land surface scheme by MIKE SHE for the FIFE  
37 area, while HIRHAM in standard configuration is used for the remaining model area.

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44 tables.  
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59 **Keywords:** Land surface – atmosphere interactions, land surface processes, regional climate modelling,  
60 evapotranspiration, groundwater  
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38 **Introduction**

139 The hydrological cycle has traditionally been studied in two separate parts. The climate and meteorological  
2  
340 communities have developed models describing the atmospheric part with redistribution of energy and  
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41 moisture, while hydrologists have developed models for the terrestrial part including processes for river, soil  
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62 moisture and groundwater (Shelton 2009). As a result hydrological impact studies of climate change have  
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843 generally been carried out by simply forcing hydrological models with output from climate models (e.g.  
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944 Graham et al. 2007; van Roosmalen et al. 2007). In this type of approach, feedback from the terrestrial part  
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1145 of the hydrological cycle to the atmosphere is neglected, which may result in significant errors (Seneviratne  
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1346 et al. 2010).

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147 In modelling the land surface energy balance, the representation of the soil moisture in particular is found to  
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1648 be crucial (Sellers and Hall 1992). Significant errors may arise if the spatial variability is not included, e.g.  
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1849 by use of simple averaging (Wood 1997). The spatial variability of soil moisture can be modelled in different  
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1950 ways. Giorgi and Avissar (1997) provided a description of different approaches of including surface  
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2151 heterogeneities in atmospheric models. In Kollet and Maxwell (2008) the land surface influence on surface  
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2352 fluxes is further investigated documenting a distinct correlation between groundwater depth and  
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2453 evapotranspiration.

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264 The effect of the feedback from soil moisture and land surface processes on the atmosphere has been studied  
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2855 in a range of studies. Miguez-Macho et al. (2007) showed that the inclusion of groundwater can lead to  
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2956 substantially wetter soils in some valley and coastal regions using the RAMS-hydro which is a non-  
30  
3157 hydrostatic regional climate model including a groundwater component. Using the same model setup, Anyah  
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3358 et al. (2008) showed that for regions where the groundwater table produces wetter soils a direct improvement  
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3459 in the reproduction of evapotranspiration is seen for dryer areas in North America. The influence of land  
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3660 surface temperatures and spatio-temporal soil moisture distribution on the simulation of sensible and latent  
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3861 heat fluxes is highlighted by Zeng et al. (2003) using the RegCM2 regional climate model over Eastern  
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3962 China.

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4163 To account for the feedback between the land surface and the atmosphere coupled hydrological-atmospheric  
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4364 models have emerged. Yuan et al. (2008) implemented a simple groundwater model into the regional climate  
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4465 model RegCM3. They found that the dynamical groundwater table affected the surface fluxes and hereby the  
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4666 boundary layers, the location of precipitation and the wind. Maxwell et al. (2011) coupled ParFlow to the  
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4867 Advanced Weather and Research model WRF (Skamarock et al. 2008) and highlighted the influence of soil  
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4968 moisture in predicting the boundary wind layer. Overgaard et al. (2007) investigated the coupling of MIKE  
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5169 SHE (Graham and Butts 2005) to ARPS (Xue et al. 2000; 2001). In a hypothetical land use change study  
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5370 they found significant differences between the results of the coupled model system and the traditional one-  
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5471 way approach, where MIKE SHE was forced by ARPS. York et al. (2002) illustrated how a future change in  
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5672 groundwater may affect the future evapotranspiration; e.g. long continuous drying, like the 1930 dust bowl,  
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5873 might lower the groundwater table making less water available for evapotranspiration. York et al. (2002) also  
59  
6074 found that groundwater dynamics had to be considered for proper long term simulation of droughts.

75 Among regional climate models (RCMs) a general bias related to the representation of surface processes was  
176 found within the ENSEMBLES project (van der Linden and Mitchell 2009). In simulations forced by the  
277 ERA-40 reanalysis many models showed an increased warm bias with increasing monthly mean  
478 temperatures (Boberg and Christensen 2012; Christensen et al. 2008). A similar warm bias was reported  
579 within the PRUDENCE project (Jacob et al. 2007). Christensen et al. (2008) explained this bias in terms of  
780 the simulated soils in the warm dry summer months which become too dry and thereby created  
981 unrealistically high sensible heat fluxes. They argued that this is because of the simplistic land surface  
1082 representation in many RCMs. Arguably, coupling the RCMs to a hydrological model may reduce the  
1283 temperature bias as the hydrological model provides a comprehensive and perhaps more accurate surface  
1484 flux calculation than the generally simpler schemes typically embedded in a RCM.  
1585 The objectives of the present study are; 1) to analyse the importance of including groundwater dynamics in  
1786 the estimation of land surface – atmosphere feedbacks, and; 2) to evaluate potential benefits and challenges  
1987 in the hydrology-land surface response when applying a fully coupled climate-hydrological model with an  
2088 advanced land surface flux scheme. The FIFE area in Kansas, USA, is used as a test case for evaluating the  
2289 effects of replacing the land surface scheme in HIRHAM regional climate model with a high resolution  
2490 MIKE SHE based hydrology and land surface model including groundwater.

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## 28 **Methodology**

### 29 *Observations /study area*

3194 During the First International Satellite Land Surface Climatological Project (ISLSCP) Field Experiment  
3295 (FIFE) (Sellers et al. 1992) an area of 15 x 15 km<sup>2</sup> near Manhattan in Kansas, USA, was intensively  
3496 monitored. The FIFE data set consists of a high number of meteorological station, soil and vegetation data  
3597 and is therefore ideal for testing land surface models (Sellers et al. 1992). The FIFE land surface can be  
3798 characterized as tallgrass prairie at an elevation between 350 – 450 m.a.s. The data used in the present study  
3999 is monitored from May 26<sup>th</sup> to October 16<sup>th</sup> in 1987, with four intensive field campaigns of roughly two  
4100 weeks duration (Betts and Ball, 1998). The soils are all either silty loam or silty clay loam (Huemmrich and  
4201 Levine 1994; Kanemasu 1994). Below the soils there are layers of limestone and shale. Since FIFE is defined  
44102 as a square and not a hydrological catchment, measurements of total discharge were not applicable.  
4603 Discharge is therefore only measured in the 12 km<sup>2</sup> King’s Creek catchment at 15 min intervals, when flow  
4804 rates exceeded 3x10<sup>-4</sup> m<sup>3</sup>/s (Wood 1994).

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### 51 *Models*

52107 MIKE SHE (Graham and Butts 2005) is an integrated distributed numerical modelling system. In this study  
54108 the model is configured using the modules for overland flow (2D diffusive wave), unsaturated flow (1D  
55109 Richards’ equation), saturated groundwater flow (3D Darcy equation) and evapotranspiration (two-layer  
57110 Shuttleworth and Wallace scheme; Overgaard 2005).

59111 HIRHAM is a regional climate model (Christensen et al. 2006). It consists of the dynamical core of the  
61112 synoptic scale weather forecast model HIRLAM (Undén et al. 2002) where the physical parameterisation

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113 schemes are replaced by those from the global circulation model ECHAM (Roeckner et al. 2003). The land  
114 surface model has five layers for calculation of soil temperature and the water budget is formulated with four  
115 reservoirs: snow intercepted by canopy, snow at surface, rainwater intercepted by canopy and soil water.  
116 For the coupled simulations MIKE SHE and HIRHAM are linked using the OpenMI software (Open  
117 Modelling Interface) (Gregersen et al. 2007) facilitating data transfer across the operational platforms of  
118 Windows and a Linux high performance parallel computation system (HPC). The coupling approach is  
119 described in Butts et al. (2014). OpenMI also handles the data exchange timing, variable definitions, unit  
120 conversions and spatial interpolation. The coupling is performed with each model operating at dedicated  
121 spatial scales: The RCM covers about half of the USA (Fig. 1), while the hydrological model only covers the  
122 15 x 15 km<sup>2</sup> FIFE area corresponding roughly to a single RCM grid. Outside the FIFE area, the land surface  
123 processes in the coupled model are based on the HIRHAM scheme. Data are exchanged between the two  
124 models every hour.

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#### 126 *Hydrological model parameterisation and data processing*

127 The MIKE SHE setup in this study is an extension of the setup described in Rasmussen et al. (2012a), where  
128 modules for saturated groundwater flow and streamflow are now included. Parameterisation of the  
129 unsaturated zone is taken directly from field data from FIFE and literature (Rasmussen et al. 2012a). In  
130 contrast, no direct data are available for parameterisation of the saturated zone and streams. Therefore, these  
131 modules are parameterised by a combination of default or literature values or literature and calibration.  
132 Calibration has been performed against discharge measurements of King's Creek (Fig. 1).

133 When running MIKE SHE in uncoupled mode the atmospheric driving data are station based from the FIFE  
134 data base consisting of ten meteorological stations with half-hourly collection (Dabberdt 1994). The  
135 classification of vegetation is based on Davis et al. (1992). The classification includes a combination of  
136 either burned or unburned and either upland, bottomland, moderate slope or steep slope. Vertically, the soils  
137 are parameterised based on the five soil profile analyses in the FIFE data base (Huemmrich and Levine 1994;  
138 Kanemasu 1994). Horizontally, the classification is based on the soil map provided in the FIFE data base,  
139 where each type has been related to one of the five soil profile, a mean soil profile or a coarser texture  
140 unknown profile (Rasmussen et al. 2012a). The unsaturated zone is initialized with soil moisture at  
141 equilibrium pressure; i.e. field capacity at the surface and full saturation at the top of the groundwater table.  
142 Layers of near-horizontal and relatively impermeable shale and thin limestone layers are underlying the  
143 surface soil (Davis et al. 1992). The hydraulic conductivity of the limestone is in the range 10<sup>-8</sup> – 10<sup>-3</sup> m/s  
144 (Macpherson 1996). The shale and limestone layers are assumed to be more fractured and porous near the  
145 surface, because of weathering. Therefore, the transport of water is assumed to only be important in the top  
146 of the shale and limestone layers. The aquifer is represented by a 2D model with the impermeable bed  
147 located at 10 m below the surface and a calibrated hydraulic conductivity of 5 x 10<sup>-6</sup> m/s. Rapid localised  
148 run-off in sub-grid scale creeks and gullies on the hills and slope are represented conceptually as drains. The  
149 drain level is 1 m below the surface and the drain constant is calibrated to 10<sup>-5</sup> s<sup>-1</sup>. The drains are activated,

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150 when the groundwater table exceeds this level, and the water is then routed to the nearest stream based on the  
151 topography.

152 The locations of streams are derived from a digital elevation model at a 25 m grid (Strebel et al. 1994). In the  
153 King's Creek catchment streams have been inserted in the model for reaches with catchments larger than 1.5  
154 km<sup>2</sup>, while this limit has been set to 10 km<sup>2</sup> for the remaining part of the FIFE area. Idealised V-shape cross  
155 sections are defined along the streams and interaction with the aquifer is only controlled by the hydraulic  
156 conductivity of the aquifer.

157 The location of the groundwater table is unknown. The initial groundwater head conditions for the FIFE  
158 observation period have therefore been simulated by running the model for a sufficiently long warm-up  
159 period so that the given initial conditions do not influence the simulation results. This was achieved by three  
160 looped simulations of the period January 1<sup>st</sup> 1985 to December 31<sup>st</sup> 1987, each time saving the groundwater  
161 heads which were then used as an initial condition for the following run.

### 162 163 *HIRHAM setup*

164 HIRHAM (version 5; Christensen et al. 2006) is run over a domain covering the central US, forced by ERA-  
165 40 reanalysis data (Uppala et al. 2005). The domain is 122 by 122 cells with a resolution of 0.125 degrees  
166 and 31 vertical levels (Fig. 1). HIRHAM uses a rotated longitude/latitude model grid. The origin in this setup  
167 is located at FIFE Lon. -96.5 Lat. 39.0. The HIRHAM simulation is started on January 1<sup>st</sup> 1987, yielding five  
168 month of spin up. In the coupling HIRHAM is started from an uncoupled restart file on May 1<sup>st</sup>.

### 169 170 *Experimental setup*

171 Five model runs were performed:

- 172 • Run "MSHE-UZ-OBS" is taken from Rasmussen et al. (2012a). This MIKE SHE run represents a  
173 distributed 1D unsaturated zone (UZ) setup with a uniform stationary groundwater table at 3 m depth  
174 and distributed meteorological forcings, soils, vegetation and overland flow. The grid resolution is  
175 60 m. The simulation starts at May 1<sup>st</sup> 1987.
- 176 • Run "MSHE-SZ-OBS" includes streamflow modelling and the saturated zone (SZ) as a single  
177 aquifer with uniform properties. The simulation is started at May 1<sup>st</sup> 1987. The initial groundwater  
178 table is taken from the final warm up run at May 1<sup>st</sup> 1987.
- 179 • Run "MSHE-SZ-HH" is identical to MSHE-SZ-OBS but the simulated precipitation input from  
180 HIRHAM-STD (see below) is used. This allows a direct comparison between evapotranspiration  
181 schemes based on the same input.
- 182 • Run "HIRHAM-STD" is a HIRHAM simulation with its own land surface model. The simulation is  
183 started on January 1<sup>st</sup> 1987.
- 184 • Run "HIRHAM-MSHE" is a coupled run of HIRHAM and MIKE SHE (including groundwater and  
185 therefore coupling from the groundwater to the atmosphere over the FIFE area). The simulation is  
186 started May 1<sup>st</sup> 1987 from an uncoupled restart file of the HIRHAM-STD run and groundwater table  
187 as for MSHE-SZ-OBS and MSHE-SZ-HH.



## 189 **Results**

### 190 *Groundwater conditions and the influence of evapotranspiration and soil moisture*

191 A direct calibration of the groundwater model was not possible, since there are no observed groundwater  
 192 head data. The data providing most relevant information on groundwater conditions are therefore the  
 193 baseflow recession parts of the discharge hydrograph at King's Creek (Fig. 2). The location of King's Creek  
 194 and the discharge station is seen in Figure 1. The simulated discharge corresponds well to the temporal  
 195 pattern and magnitude of the observations and the recession parts of the simulated hydrograph have largely  
 196 similar shape as in the observed hydrograph. The similar shape of the hydrograph recessions indicates that  
 197 the baseflow conditions, and hence the overall stream-aquifer interactions, are represented by the model. A  
 198 perfect match in the timing of peaks cannot be expected in the warm up period (before May 1<sup>st</sup> 1987) as the  
 199 precipitation input is not recorded within the catchment. In the FIFE period only one peak occurs at the end  
 200 of May to mid-June.

201 Another indication of a plausible representation of groundwater by the model is the depth to the groundwater  
 202 table, which is expected to be small in the bottom of the valleys and larger at the hill tops. Wood (1997)  
 203 estimated the mean water table depth, based on the soil-topographic index of TOPLATS (Famiglietti and  
 204 Wood 1994), to be 3.5 m in dry conditions and 2.0 m in wet conditions for the King's Creek catchment. The  
 205 simulated depth to the groundwater table at the beginning of the FIFE period is seen in Figure 1 where the  
 206 mean depth to the groundwater is 3.6 m for the whole FIFE area and 4.3 m for the King's Creek catchment.  
 207 The simulated spatial distribution of depth to the groundwater is similar to the distribution of the soil-  
 208 topographic index by Famiglietti and Wood (1994).

209 To illustrate the direct influence of the groundwater table on evapotranspiration, Figure 3 shows the  
 210 simulated groundwater depths and evapotranspiration for three grid cells with the same vegetation, soil type  
 211 and meteorological forcing. The three cells are selected as one of the highest and one of the lowest depths to  
 212 groundwater table and one in between. The two cells with ~2 and 8 m depth to the groundwater table have  
 213 similar evapotranspiration, while the cell with shallow groundwater table (0-0.5 m below the surface) has a  
 214 much larger evapotranspiration.

215 Figure 4 shows the depth to the groundwater table plotted against evapotranspiration on August 1<sup>st</sup> for cells  
 216 of two selected soil types for the MSHE-SZ-OBS simulation. The greatest spread in evapotranspiration, in  
 217 cells with groundwater depth greater than 2 m, is seen among the different classes of soil types; exemplified  
 218 by the two shown soil types. The spread among vegetation types and meteorological stations is similar (not  
 219 shown). The largest variation in the relationship between evapotranspiration and depth to groundwater is  
 220 found among these soils for cells with groundwater depth less than 2 m. Each point on Figure 4 has been  
 221 coloured according to meteorological forcing. The different meteorological stations result in different levels  
 222 of evapotranspiration at groundwater depth greater than 2 m. Depending on the soil type, the depth where the  
 223 groundwater table becomes unimportant varies from 1 to 3 m (Soil type Silty loam to Silt - not shown). The  
 224 results are to some extent influenced by the model setup: (i) as the drainage depth is located at 1 m below the  
 225 surface in all grids, except at the bottomland vegetation in the low lying river valleys, the model may not be

226 able to correctly simulate shallow groundwater table of less than 1 m depth; and (ii) the vertical  
227 discretization in the unsaturated zone has some effect on the calculated groundwater table.  
228 Observed and simulated areal mean evapotranspiration and soil moisture changes in the unsaturated zone  
229 during the FIFE period are shown in Figure 5. The simulated evapotranspiration by MSHE-UZ-OBS and  
230 MSHE-SZ-OBS are largely similar, except for the dry period from mid-July to beginning of August, where  
231 MSHE-SZ-OBS has a higher evapotranspiration than MSHE-UZ-OBS. Also, the MSHE-UZ-OBS and  
232 MSHE-SZ-OBS simulations reproduce the observed evapotranspiration levels and temporal patterns whereas  
233 the MSHE-SZ-HH evapotranspiration is distinctly different. The soil moisture as simulated by MSHE-SZ-  
234 OBS is continuously drier than MSHE-UZ-OBS but otherwise comparable (Fig. 5). Most of this difference  
235 can be explained by the differing groundwater table forming the lower boundary condition for initialisation  
236 of the unsaturated zone with equilibrium pressure condition, which results in different initial soil moisture  
237 contents.

#### 238 239 *HIRHAM simulations and the effect of the land surface*

240 The influence of land surface scheme on the simulation of six meteorological variables is shown in figure 6.  
241 This compares the output from HIRHAM-STD and HIRHAM-MSHE with observations for the RCM grid  
242 cell covering 72 % of the FIFE area. For the entire period there is generally a reasonable match between  
243 simulations and observations. Some exceptions include overestimation of air temperature and discrepancies  
244 in incoming shortwave radiation. These are affected by differences in cloud cover between simulations and  
245 observations where spatial scale issues and cell averaging have a large effect. For the single HIRHAM cell  
246 over the FIFE area precipitation does not occur on the same days as observed and is generally over-  
247 estimated, most notably for the HIRHAM-MSHE simulation. From May 1<sup>st</sup> to Oct 16<sup>th</sup> the observed  
248 precipitation is 495 mm whereas the HIRHAM-STD and HIRHAM-MSHE levels are 684 and 1523 mm  
249 respectively. The largest HIRHAM-MSHE overestimations occur during the first half of the FIFE period.  
250 This geographical region is subject to highly dynamic convective precipitation with a high degree of  
251 variability on spatial scales compared to the HIRHAM grid cell resolution of the present study. To illustrate  
252 this, precipitation output from the 5 RCM cells surrounding the FIFE grid cell (11x11 cell output size; 150  
253 km x 150 km area) are extracted and these vary between 420-1310 mm for HIRHAM-STD and 542-1777  
254 mm for HIRHAM-MSHE.

255 The comparison between the uncoupled HIRHAM-STD and the coupled HIRHAM-MSHE shows that the  
256 difference between the two models is negligible for many meteorological variables, such as surface air  
257 pressure. This is not surprising, since the coupling is very localized, i.e. MIKE SHE is only replacing the  
258 HIRHAM-STD scheme in the local 15 x 15 km<sup>2</sup> FIFE area. However, for some variables such as air  
259 temperature and precipitation there are notable differences (Fig. 6).

#### 260 261 *Effect of meteorological forcing and land surface scheme on evapotranspiration and soil moisture*

262 The MSHE-SZ-HH simulated evapotranspiration show a distinctly different dynamics than the observation  
263 driven simulations (Fig. 5). This clearly shows the influence of precipitation on evapotranspiration. The

264 simulated daily variations in evapotranspiration do not appear to match and the main variations appear  
265 related to precipitation events (Fig. 6). For MSHE-SZ-HH the highest precipitation events occur around 27  
266 May – 2 June, 4-12 July and in mid-August. The general evapotranspiration level, and total period budget, is  
267 however reproduced with a sum of 531 mm compared to the observed 540 mm (Table 1). Similarly to the  
268 simulated evapotranspiration, MSHE-SZ-HH reflects positive soil moisture changes temporally related to  
269 precipitation events in, especially, early July and early August.

270 Figure 7 shows simulated evapotranspiration and soil moisture change of the unsaturated zone by MSHE-  
271 SZ-OBS, HIRHAM-STD and HIRHAM-MSHE. Firstly, it should be noted that the nature of the simulations  
272 resulted in substantial differences in precipitation input complicating the analysis (Table 1) and especially  
273 the HIRHAM-MSHE run is an outlier with approximately three times the observed precipitation during the  
274 period. These differences are also reflected in the evapotranspiration levels especially for June-July.

275 The differences in the MIKE SHE and HIRHAM land surface schemes is seen to have a substantial effect on  
276 evapotranspiration and soil moisture (Fig. 7 and Table 1): The precipitation input for MSHE-SZ-HH and  
277 HIRHAM-STD is the same and differences are therefore due to energy flux schemes and the complexity in  
278 hydrological processes, and in particular groundwater. Comparing the MSHE-SZ-HH and HIRHAM-STD  
279 evapotranspiration, the latter shows a more dynamical behaviour and general higher levels (531 and 586 mm  
280 respectively, table 1). In the dry period, from 13 July to 12 August, MSHE-SZ-HH more closely resembles  
281 the observations than HIRHAM-STD, where the evapotranspiration is generally higher. For soil moisture  
282 MSHE-SZ-HH and HIRHAM-STD are more similar except in the dry mid-summer period where HIRHAM-  
283 STD is depleted at a faster rate than MSHE-SZ-HH which also resembles the observations by having the  
284 same gradient. The high precipitation input in the HIRHAM-MSHE run is also reflected in overall higher  
285 evapotranspiration levels as well as peaks and the soil moisture change patterns resemble the HIRHAM-STD  
286 with variations related to differences in precipitation. The precipitation in HIRHAM-MSHE is around twice  
287 as high as in the HIRHAM-STD simulation, the runoff is nearly three times as high (Table 1) whereas the  
288 evapotranspiration is 1.3 mm/d higher. Thus, a high intensity precipitation event causes increased runoff due  
289 to exceedance of infiltration capacity, whereas the soil moisture and evapotranspiration is less affected.

290 In HIRHAM-MSHE the general levels of simulated evapotranspiration are higher than the MIKE SHE based  
291 simulations and equal to the HIRHAM-STD simulation at all groundwater depths as seen for 1 August in  
292 figure 8a. The MSHE-SZ-OBS and MSHE-SZ-HH evapotranspiration levels are generally comparable  
293 although the latter is higher for depths to the groundwater lower than app. 1.75 m and vice versa. The  
294 groundwater depth at which there is an influence on the evapotranspiration diminishes at levels of 0.5-3 m,  
295 and evapotranspiration reaches levels of 6-11 mm/day for the highest groundwater levels. All three  
296 simulations with distributed groundwater have a peak in groundwater depths at the intervals of 0.5-0.75 m or  
297 0.75-1 m and there is a tendency for a positive correlation with precipitation amount input and share of lower  
298 groundwater depth grid cells (Fig. 8b).

## 300 Discussion

301 The present study shows results from simulation with two different land surface schemes, varying  
302 groundwater conditions and varying precipitation input. It is shown that enhancements in the representation  
303 of groundwater improve the simulation of evapotranspiration (Fig. 5) and, not unexpectedly, that  
304 precipitation amounts as well as timing is equally important to reproduce evapotranspiration (Fig. 5 and 7).  
305 However, the 178 mm difference between observed and simulated (HIRHAM-STD) precipitation input  
306 results in negligible differences in evapotranspiration where instead the unsaturated storage and river  
307 drainage is affected. Evidence of the potentially added value of coupling an RCM to a hydrological model,  
308 especially in dry periods where evapotranspiration substantially exceeds the net precipitation is seen. In the  
309 dry period within the FIFE experiment (mainly July and start August) HIRHAM-STD overestimates the  
310 evapotranspiration, implying that the soil moisture is not a limiting factor. In contrast, MSHE-SZ-HH, using  
311 the same precipitation input, simulates a drying of the soil and a reduction in the evapotranspiration (Fig. 7).  
312 The MSHE-SZ-OBS simulation, including groundwater, shows how evapotranspiration is high during the  
313 dry period for the cells with shallow groundwater table (Fig. 3). The MSHE-UZ-OBS simulation with  
314 uniform groundwater table at 3 m does not have shallow groundwater in the valleys and cannot maintain  
315 evapotranspiration during the dry out period (Fig. 5), while it is able to represent evapotranspiration in areas  
316 with a groundwater depth >2 m (Fig. 8a). The simulations suggest that the area, where groundwater is less  
317 than 2 m deep, represents 35 % of the total FIFE area at the end of the dry out period. Resolving the  
318 groundwater conditions adequately requires a hydrological model with fine resolution, depending on the  
319 topography of the catchment and especially the width of the river valleys. Typically, the soil and vegetation  
320 may differ in the valleys compared to the surroundings. Therefore, a finer spatial resolution of the valleys  
321 may be required to simulate the effects of different soils (like Fig. 4). In our case grid sizes in the order of  
322 60-240 m are required (Rasmussen et al. 2012a). If, however, the effect of shallow groundwater is  
323 disregarded, the hydrological model can be run at much coarser resolution as long as the variation in  
324 vegetation and soil types are preserved (Rasmussen et al. 2012a). The mosaic approach used here for  
325 modelling the land surface with a high resolution hydrological model is computationally demanding.  
326 Running a coupled climate simulation with MIKE SHE at 60 m resolution for the whole HIRHAM domain is  
327 not feasible with the current computational capacity and would also require extensive validation of the MIKE  
328 SHE model. The local coupling approach allows the high resolution hydrological model to be applied only  
329 for areas of particular importance for the land surface atmosphere interaction or for areas where the  
330 hydrological effects of climate change are of interest. Computationally more efficient, statistical methods  
331 exist for handling subgrid variability in soil, vegetation, topography and groundwater depth, but they do not  
332 allow changes in groundwater table due to e.g. climate change.

333 This study found differences in daily evapotranspiration rates of ~6 mm/day in the dry period between two  
334 similar cells with groundwater tables in 0.5 and 2 m depth, respectively. Likewise, Kollet and Maxwell  
335 (2008) and Maxwell and Kollet (2008) found strong correlation between depth to groundwater and  
336 evapotranspiration at groundwater depths, which they term the critical zone. In their study of the Little  
337 Washita in Oklahoma the critical zone is between 1 and 5 m below the surface, while the corresponding  
338 depths in our results are between 1 and 3 m below the surface. Considering that the two studies are from

339 different locations with differences in soils and vegetation, model codes with different conceptualisations of  
340 unsaturated zone flows and root water uptake and that none of the studies are validated against field data, our  
341 results support the overall conclusion of Kollet and Maxwell (2008) and Maxwell and Kollet (2008). This  
342 finding emphasizes the potential of using a fully coupled system for proper climate simulations as the  
343 accumulated difference in evapotranspiration rates under more extreme hydrological conditions may be large  
344 and hence determine whether the model will be able to represent a hydraulic drought or not.

345 In relation to the generally acknowledged warm bias in warm months among RCMs over Europe  
346 (Christensen et al. 2008), our study shows that inclusion of groundwater in a hydrological model coupled to a  
347 RCM can maintain a high evapotranspiration during dry summers; and, hence, better partition the available  
348 energy between sensible and latent heat fluxes. In view of this, coupled climate-hydrological models have a  
349 potential for improving the simulations of soil surface temperature. In this study it is found that the coupled  
350 HIRHAM-MSHE model has a slightly different mean bias of the simulated 2 m air temperature compared to  
351 the default HIRHAM-STD model, but none of the two models are found to be significantly better in their  
352 overall fit to the temperature observations. For the present study however, the precipitation bias for the  
353 coupled cell is clearly affected by the coupling which leads to a different water balance for the coupled area  
354 and, hence, surface energy balance.

355 The coupled precipitation bias is likely to be caused by either: i) climate model variability as induced by the  
356 perturbation from introducing the coupling, ii) effect of land surface feedback or iii) numerical shock from  
357 overwriting HIRHAM-STD internal variable which feed into other physical equations and parameterisations.  
358 It is outside the scope of the present study to decide which combination of these, maybe in combination, is  
359 the most likely cause. Instead, the coupled simulation is simply used as an additional method of forcing the  
360 land surface calculations. Minor perturbations (or change in the model setup like the domain or resolution)  
361 are however known to cause two RCM simulations to differ even though the RCM setups otherwise are  
362 identical (Miguez-Macho et al. 2004; Rasmussen et al. 2012b; Larsen et al. 2014). Also, the region of the  
363 study in Kansas, USA, has previously shown high degrees of spatio-temporal variability related to  
364 convective precipitation (Rasmussen et al. 2012b). Another use of the same HIRHAM – MIKE SHE  
365 coupling is seen in Butts et al (2014) and Larsen et al. (2014) covering the 2500 km<sup>2</sup> Skjern River catchment  
366 in Denmark featuring multiple one-year model runs and coupling 23 HIRHAM cells in 11 km resolution. As  
367 for the present, Larsen et al. (2014) found poorer precipitation results for coupled simulations as compared to  
368 uncoupled results when assessing hourly to daily dynamics. However, longer term precipitation was  
369 improved which more recently was further emphasized for a six-year simulation where the coupled  
370 precipitation bias was significantly diminished compared to uncoupled (Larsen 2014).

## 371 372 **Conclusions**

373 Our results show that evapotranspiration is highly dependent on the depth to the groundwater, especially  
374 during dry periods, and that a correct groundwater representation therefore is important. Further, and not to  
375 much surprise, the timing in precipitation input is highly influencing the timing of evapotranspiration, which  
376 is therefore highly dependent on the input source to obtain exact daily dynamics. Longer term

377 evapotranspiration as reflected here by the 144 day simulation period is simulated equally well using a  
378 precipitation input amount of additional 36%. The HIRHAM-STD evapotranspiration compares to observed  
379 data with only an 8.5% error, but in the dry period the evapotranspiration is too high, because it is not limited  
380 by the drier soil. This could be due to the wet bias of the RCM in the preceding period but may be caused by  
381 surface parameters that are not well adjusted to the local surface conditions. By the coupling the wet bias is  
382 even higher, which further change the water and energy balance from the observed. The hydrological model  
383 in contrast shows a reduction in evapotranspiration with soil moisture in the drying period, as expected.  
384 With this study improved results were obtained by improving the detail and process range in the land surface  
385 and groundwater hydrology and energy balance. The potential of coupling an RCM and a more complex  
386 hydrology model has therefore been demonstrated under the conditions found in the central USA. Additional  
387 studies where the catchment size and the simulation period are increased would further highlight the true  
388 benefits of the coupled model setup in the present region as has already been done in a study over a Danish  
389 catchment (Larsen et al. 2014).

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**Tables**

**Tbl. 1** Water balance at FIFE over the FIFE period May 26<sup>th</sup> to October 16<sup>th</sup> 1987; accumulated mm. The difference in precipitation between HIRHAM-STD and MSHE-SZ-HH (using HIRHAM-STD precipitation as input) is due to the extraction from the central HIRHAM cell (72% of the FIFE area) and the FIFE weighted mean respectively

Simulation	Pre.	ET	UZ	SZ	Drain to River	Drain to boundary	Baseflow
Obs.	495	540	-91*	-	26	-	-
MSHE-UZ-OBS	495	508	-53	-	-	35	-
MSHE-SZ-OBS	495	542	-70	-25	36	11	1
MSHE-SZ-HH	673	531	-50	6	121	34	2
HIRHAM-STD	684	586	-68	-	207	-	-
HIRHAM-MSHE	1523	756	-74	62	620	158	2

\* Relative to May 30<sup>th</sup> or -41 mm assuming 50 mm on May 30<sup>th</sup> as for Fig 2.

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573 **Figures**

574

575 **Fig. 1** (a) HIRHAM model domain and location of the FIFE study area. (b) Depth to groundwater table,  
576 simulated at the beginning of the FIFE period May 1987 by the third warm up run. The black solid line is the  
577 King's Creek catchment with the discharge station at the black dot. Cell id labels of the three selected cells  
578 used in Fig 4 are shown to the right

579

580 **Fig. 2** Discharge of King's Creek observed and simulated by MSHE-SZ-OBS (the third warm up run)

581

582 **Fig. 3** Evapotranspiration and groundwater table at three selected grid cells of MSHE-SZ-OBS. All cells  
583 have the same vegetation type (Burned Bottomland), soil type (Florence-Benfield) and meteorological  
584 station (4439). The locations of the three grid cells are shown in Fig 1

585

586 **Fig. 4** Groundwater depth against daily mean evapotranspiration simulated by MSHE-SZ-OBS for each cell  
587 within soil type Florence-Benfield and silty loam, respectively, at August 1<sup>st</sup>. Each point is coloured  
588 according to the meteorological station used for forcing

589

590 **Fig. 5** Evapotranspiration and soil moisture change in the unsaturated zone: Observed and simulated by  
591 MSHE-UZ-OBS, MSHE-SZ-OBS and MSHE-SZ-HH. The values are mean levels over the FIFE area

592

593 **Fig. 6** Atmospheric forcing simulated by HIRHAM-STD compared to FIFE observations and HIRHAM-  
594 MSHE July 28<sup>th</sup> to August 4<sup>th</sup> for the one HIRHAM cell covering 72 % of the FIFE area (upper panels).  
595 Daily precipitation as mean over the FIFE area for the full FIFE period with HIRHAM-MSHE precipitation  
596 levels in text for two days exceeding the y-axis limit (lower panel)

597

598 **Fig. 7** Evapotranspiration and soil moisture change in the unsaturated zone simulated by MSHE-SZ-HH,  
599 HIRHAM-STD and HIRHAM-MSHE; area mean over the FIFE area

600

601 **Fig. 8** (a) Groundwater depths against daily mean evapotranspiration simulated by the five simulations at  
602 August 1<sup>st</sup>; the last day in a dry period for all three precipitation sources in the study (observations,  
603 HIRHAM-STD and HIRHAM-MSHE). Solid lines are running mean over 0.25 m intervals of depths to the  
604 groundwater. The grey diamond is the mean evapotranspiration by MSHE-UZ-OBS where the groundwater  
605 is prescribed uniform at 3 m depth. The black diamond is the mean evapotranspiration by HIRHAM-STD  
606 that does not specify the groundwater depth. (b) Percentage of grid cells within each 0.25 m interval of depth  
607 to the groundwater.

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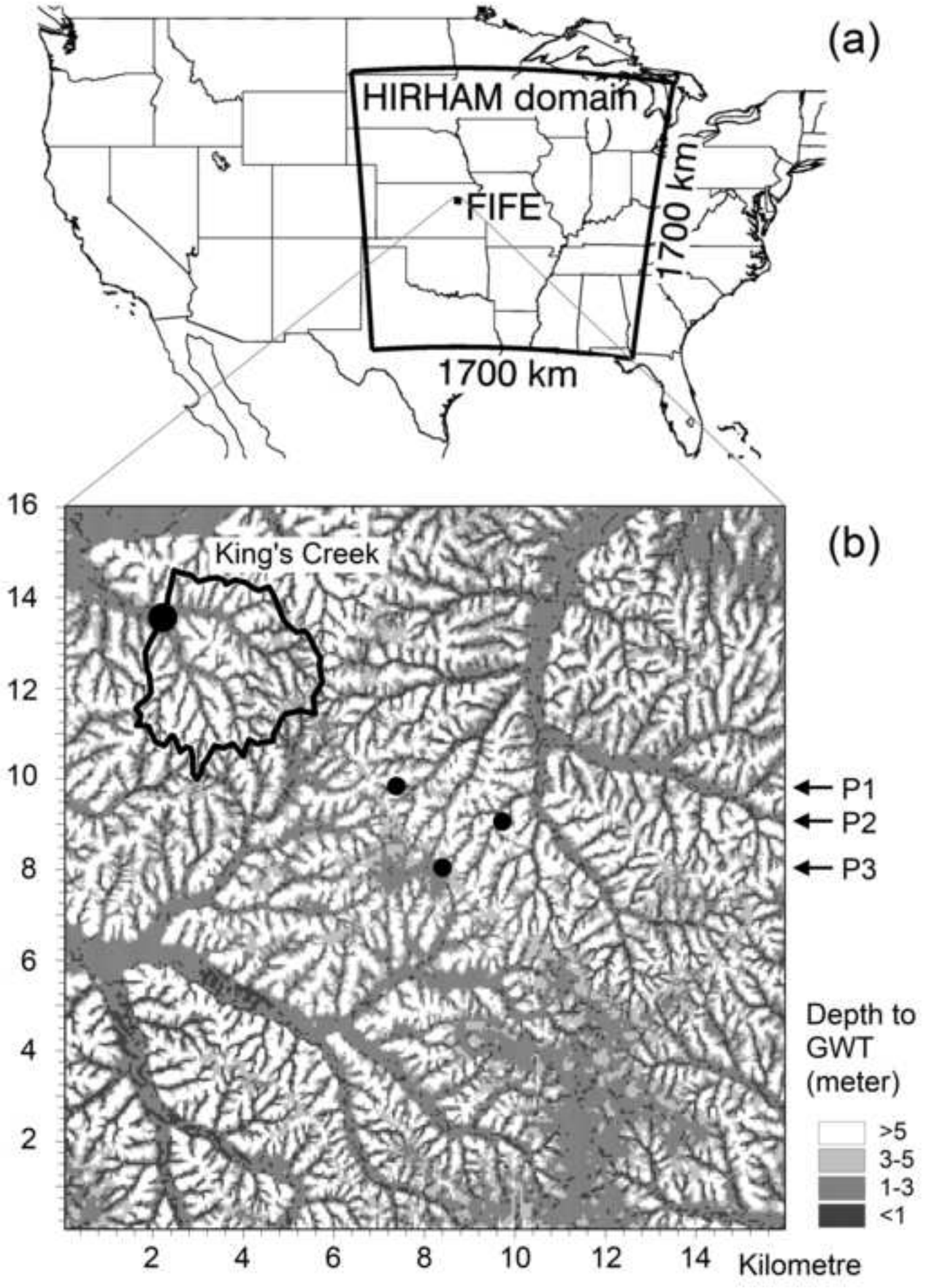
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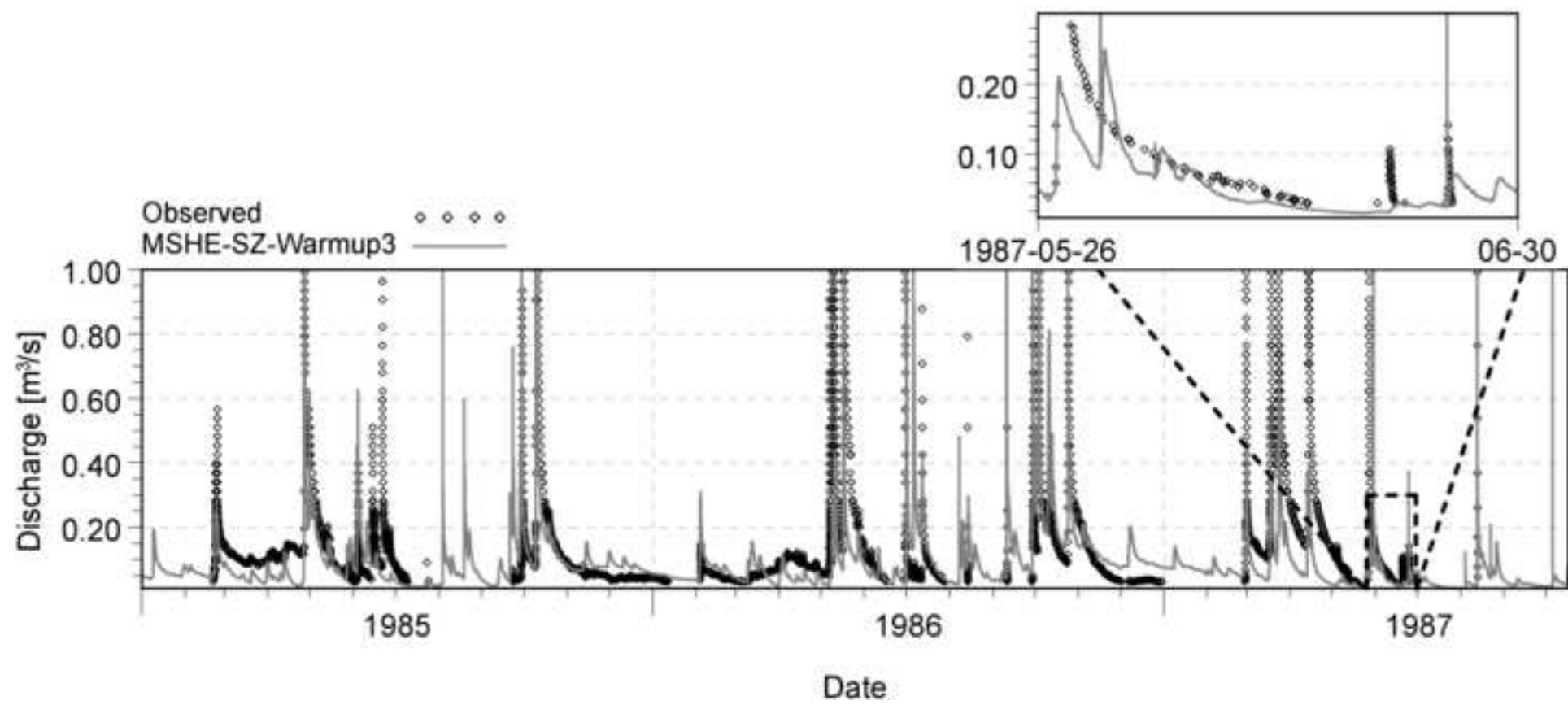
65

Figure  
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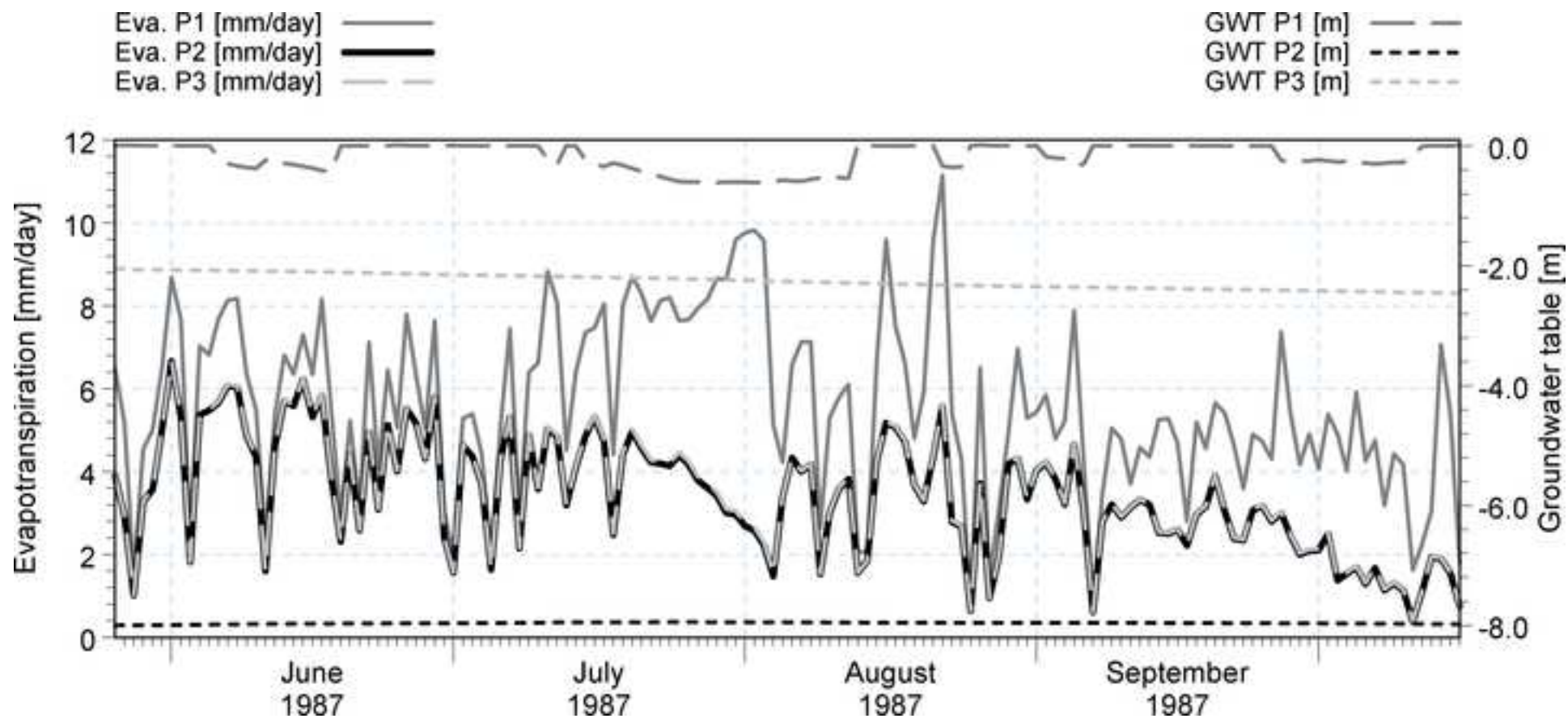
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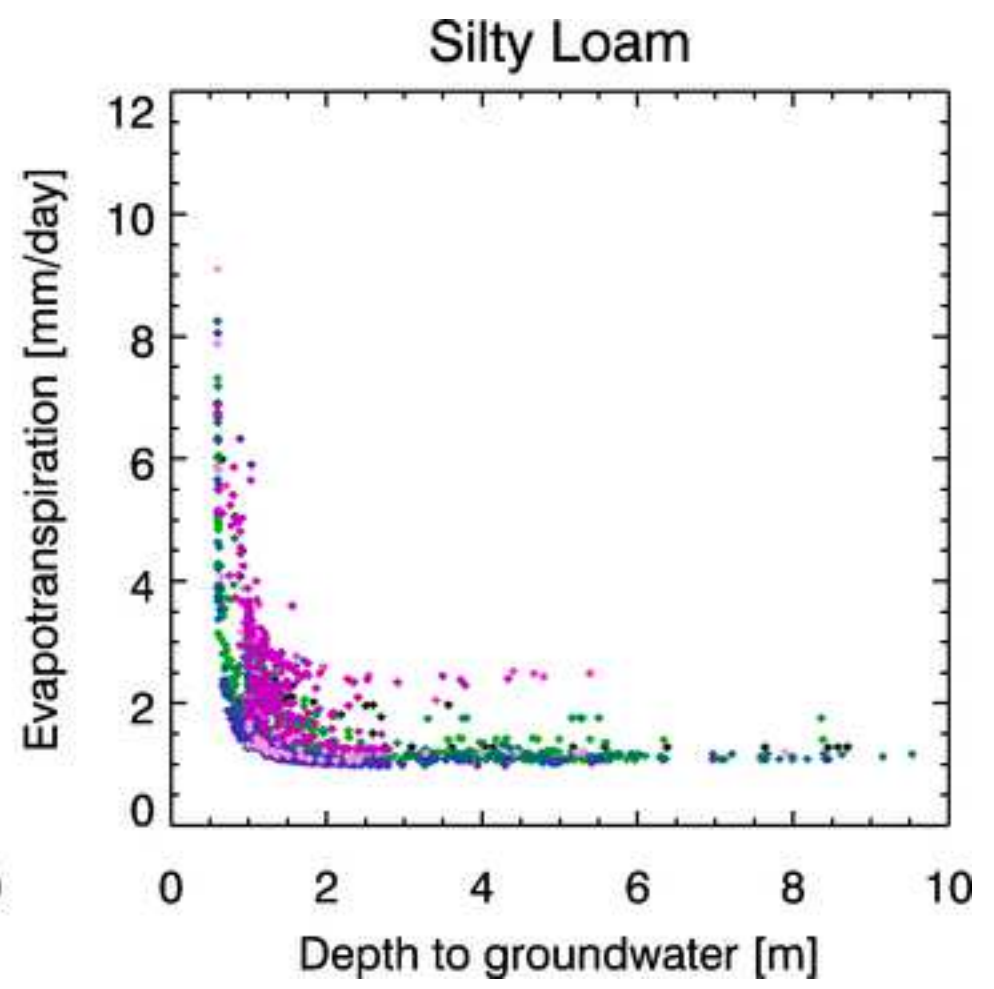
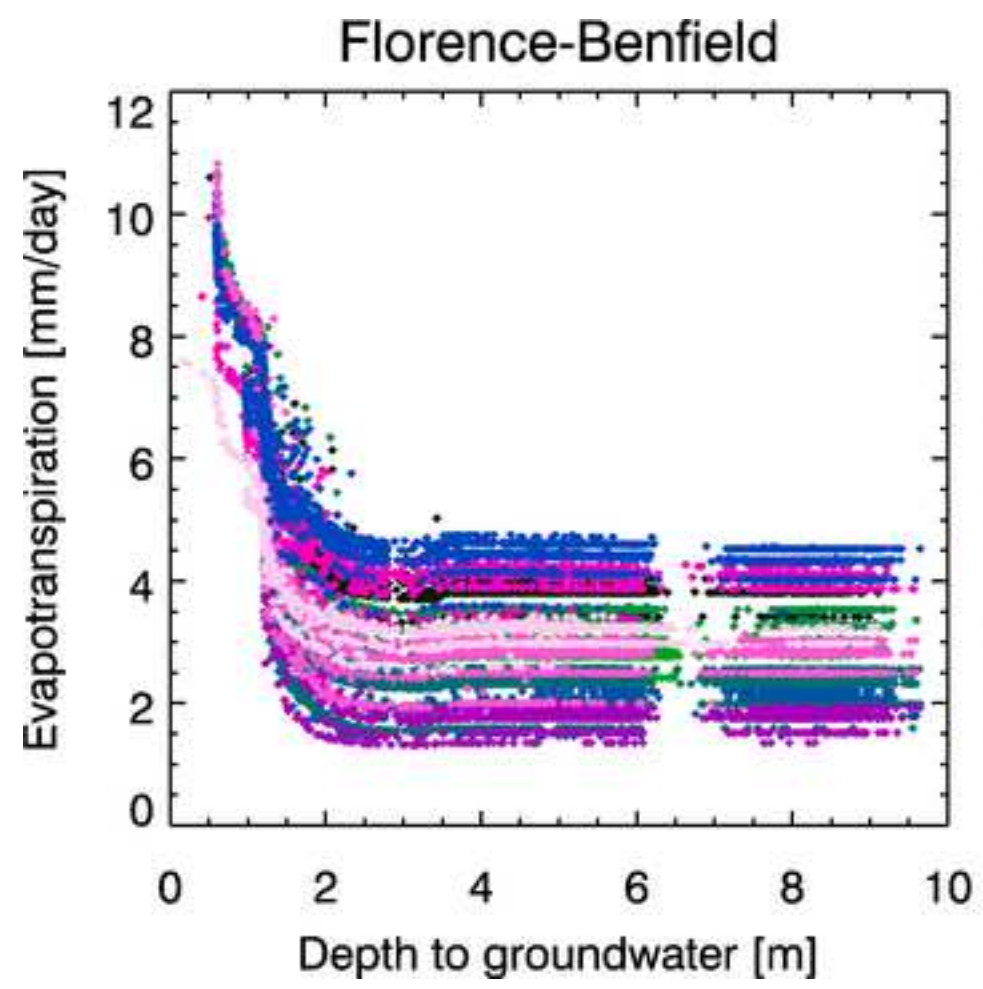
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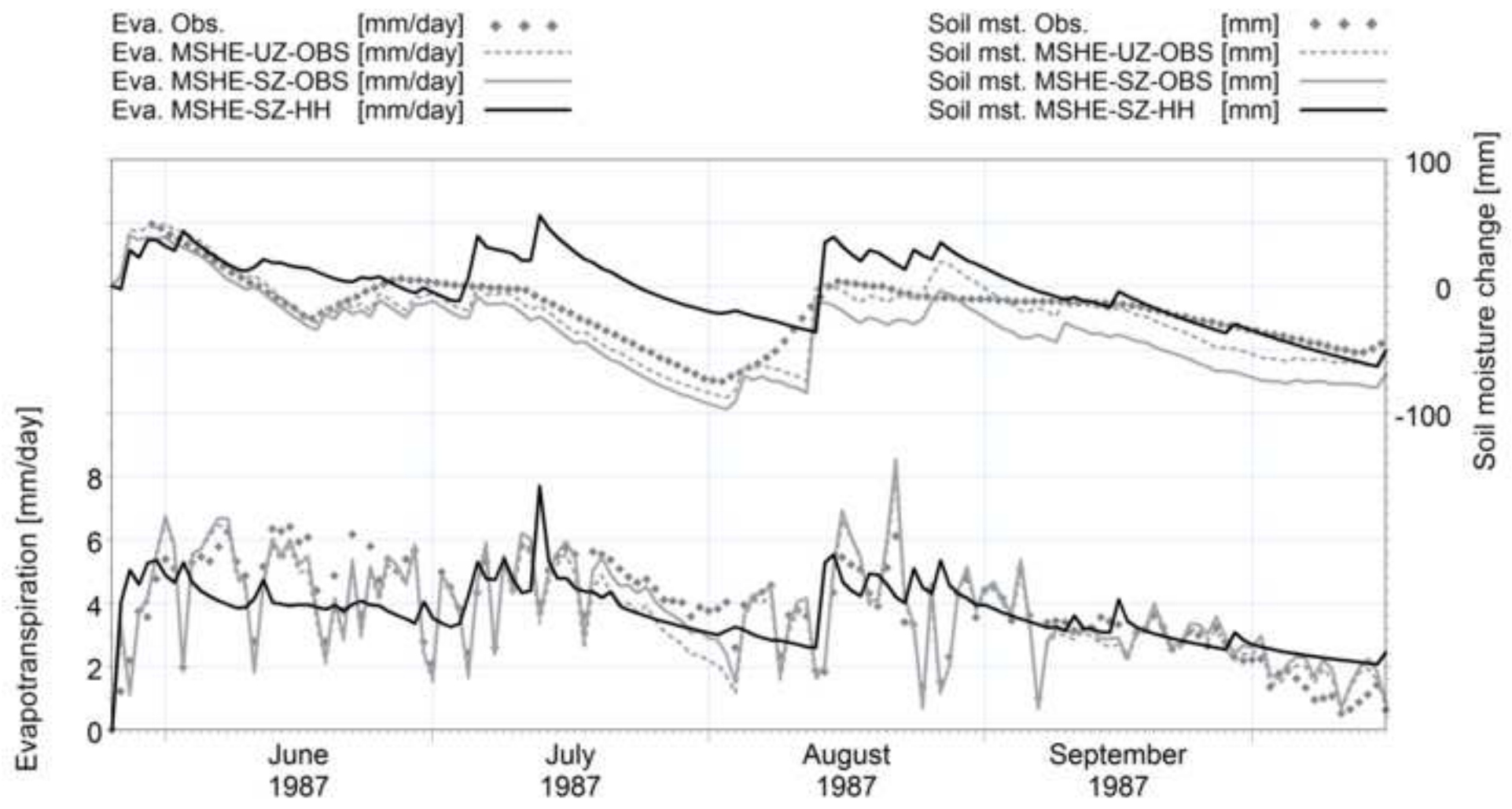
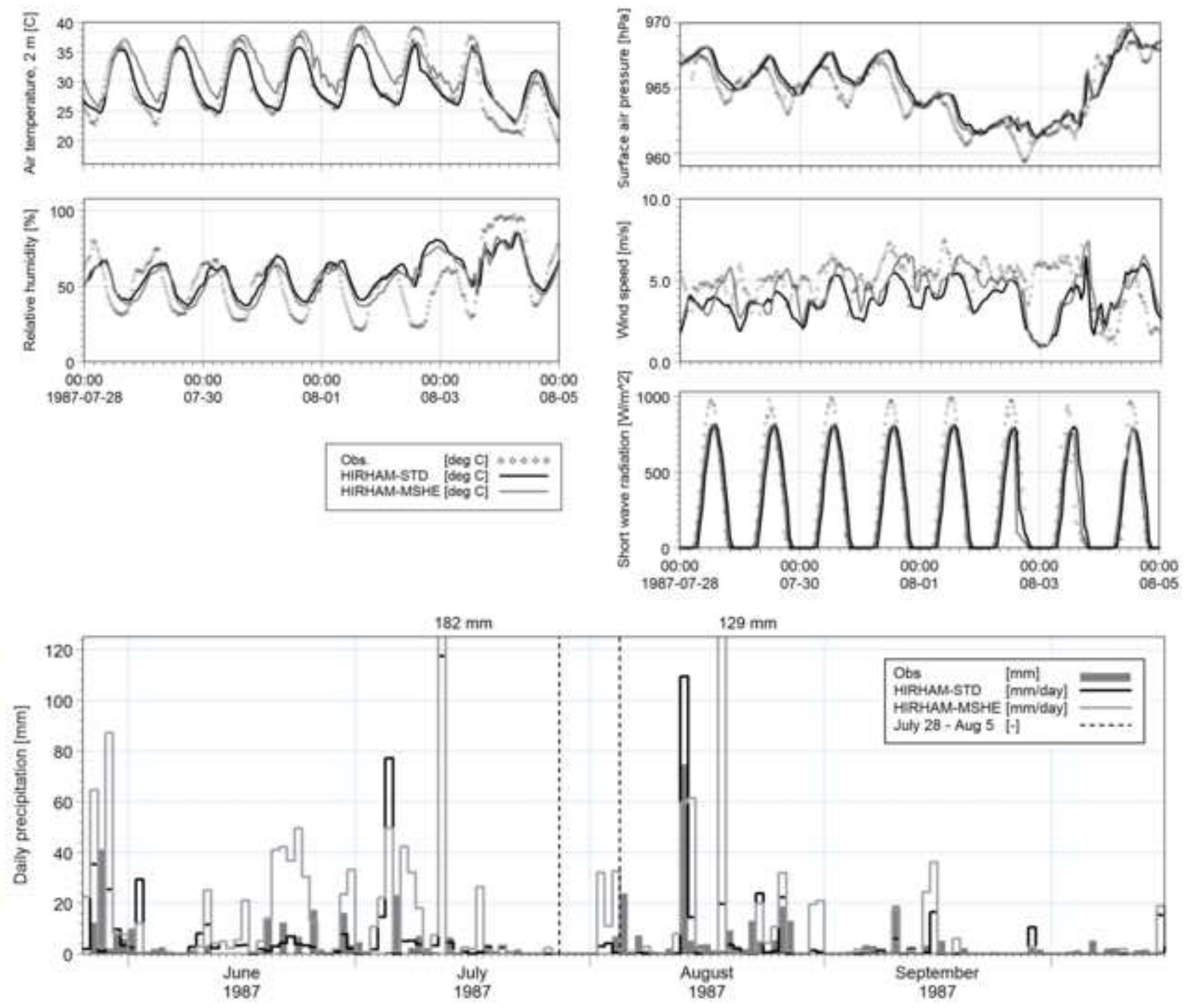




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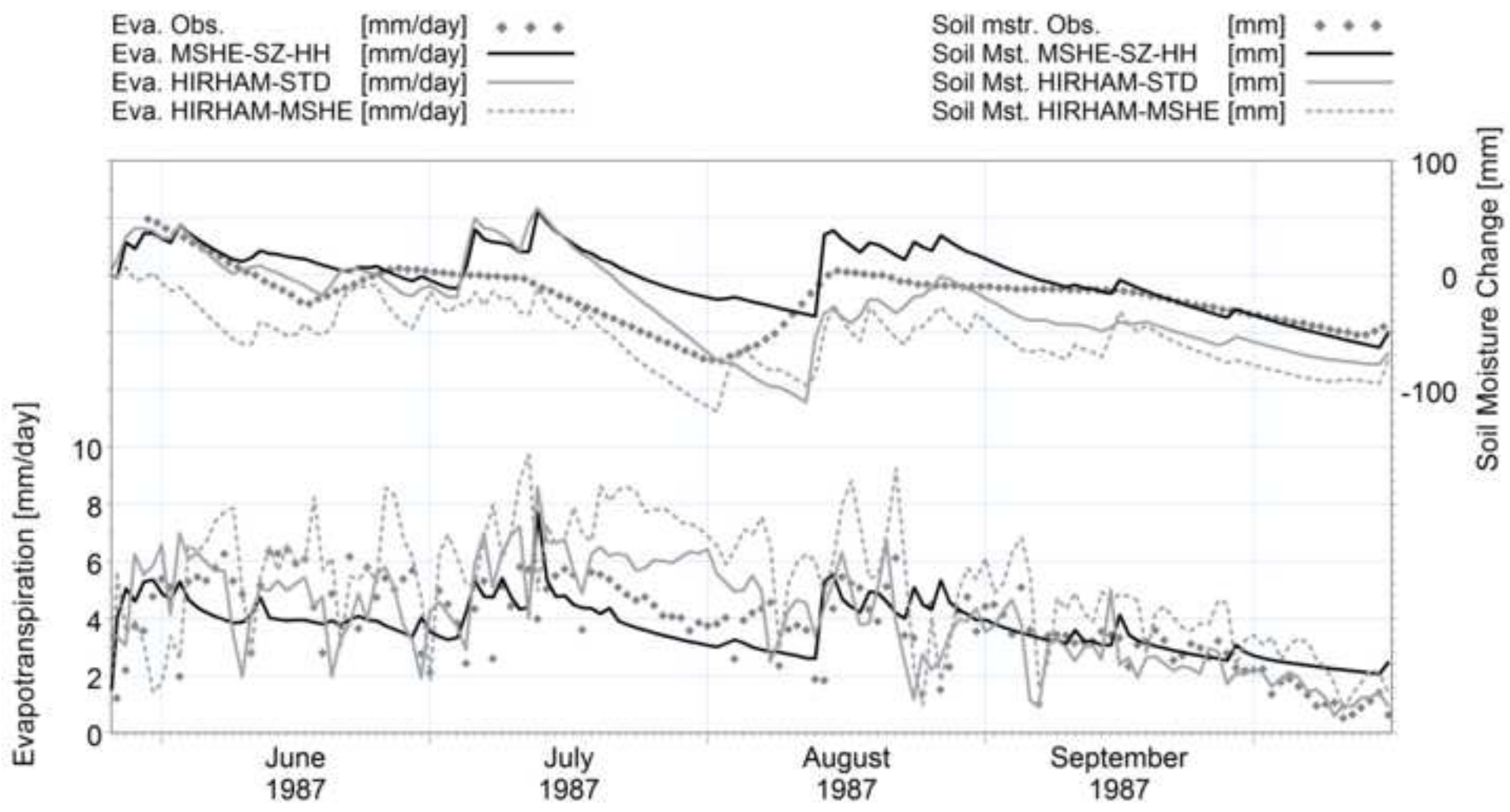


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