



1     **Modeling groundwater responses to climate change in the Prairie Pothole Region**

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6     Zhe Zhang<sup>1</sup>, Yanping Li<sup>1</sup>, Michael Barlage<sup>2</sup>, Fei Chen<sup>2</sup>, Gonzalo Miguez-Macho<sup>1</sup>, Andrew Ireson<sup>1</sup>,

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Zhenhua Li<sup>1</sup>

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<sup>1</sup>*Global Institute for Water Security, University of Saskatchewan, Saskatoon, SK, Canada*

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<sup>2</sup>*National Center for Atmospheric Research, Boulder, Colorado, USA*

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<sup>3</sup>*Nonlinear Physic Group, Faculty of Physics, Universidade de Santiago de Compostela, Galicia, Spain*

12



13 Abstract

14 Shallow groundwater in the Prairie Pothole Region (PPR) is recharged predominantly by snowmelt  
15 in the spring and may supply water for evapotranspiration through the summer/fall. This two-way  
16 exchange is underrepresented in land-surface models. Furthermore, the impacts of climate change  
17 on the groundwater recharge are uncertain. In this paper, we use a coupled land and groundwater  
18 model to investigate the hydrologic cycle of shallow groundwater in the PPR and study its response  
19 to climate change at the end of the 21st century. The results show that the model reasonably  
20 simulates the water table depth (WTD) and the timing of recharge processes, but underestimates  
21 the seasonal variation of WTD, due to mismatches of the soil types between observations and the  
22 model. The most significant change under future climate occurs in the winter, when warmer  
23 temperature changes the rain/snow partitioning, delay the time for snow accumulation/soil  
24 freezing while bring forward early melting/thawing. Such changes lead to an earlier start to a  
25 longer recharge season, but with lower recharge rates. Different signals are shown in the eastern  
26 and western PPR in the future summer, with reduced precipitation and drier soils in the east but  
27 little change in the west. The annual recharge increased by 25% and 50% in the eastern and western  
28 PPR, respectively. Additionally, we found the mean and seasonal variation of the simulated  
29 WTD are sensitive to soil properties and fine-scale soil information is needed to improve  
30 groundwater simulation on regional scale.

31

32 Keywords: Groundwater, Recharge, Climate Change, Prairie Pothole Region, Hydrologic Cycle,



33 Introduction

34 [1] Groundwater (GW) is an important source of freshwater for human beings. The domestic needs  
35 of about half of the world's population (UNESCO, 2004) and 38% of the global water demand  
36 for irrigation are provided by groundwater (Siebert et al., 2010). In the Canadian prairies, more  
37 than 30% of the population relied on groundwater in 1996 (Statistics Canada, 1996). In a more  
38 recent survey, while 90% of the municipal population is now provided by surface water sources,  
39 more than 50% of the population living in rural areas are still relying on groundwater sources in  
40 Canada (Environment Canada, 2011). In the U.S., up to 90% of water for drinking and irrigation  
41 are provided by groundwater across different parts of the country (National Research Council,  
42 2003).

43

44 [2] The groundwater flows in cold regions exhibit unique hydrological characteristics due to the  
45 hydraulic isolation, induced by seasonally frozen soil (Ireson et al., 2013). As frozen soils reduce  
46 permeability and snow accumulates during winter, the timing of groundwater recharge is  
47 controlled by the snowmelt and soil thaw period in spring. Previous work by Kelln et al. (2007)  
48 found that the timing of the recharge was associated strongly with soil thaw, rather than snowmelt,  
49 and occurred one to six weeks later than snowmelt. On the other hand, previous observations in a  
50 glacial-till site, where groundwater flow to underlying aquifer and lateral flow are small, have  
51 suggested the decline of water table during winter is related to an upward water transport to the  
52 freezing front (Remenda et al., 1996).

53

54 [3] The Prairie Pothole Region (PPR) in North America is located in a semi-arid and cold region,  
55 where evapotranspiration (ET) exceeds precipitation (PR) in summer and near-surface soil is



56 frozen in winter (Gray, 1970; Granger and Gray, 1989; Hayashi et al., 2003; Pomeroy et al., 2007;  
57 Ireson et al., 2013; Dumanski et al., 2015). The groundwater recharge to shallow aquifers in the  
58 PPR is crucially influenced by these climatic conditions and seasonal freeze-thaw processes.  
59 During the winter, snow accumulation and frozen near-surface soil prohibit infiltration. At the  
60 same time, the water table slowly declines due to a combination of upward transport to freezing  
61 front by the capillary effect and discharge to river (Ireson et al., 2013). In the early spring,  
62 snowmelt becomes the dominant component of the hydrologic cycle and the melt water runs over  
63 frozen soil, with little infiltration contributing to recharge. As the soil thaws, the increased  
64 infiltration capacity allows snowmelt recharge reaching the water table, the previously upward  
65 water movement by capillary effect move downwards, and water table rises to its maximum level.  
66 In the summer and fall, when high ET exceeds PR and desiccates the soil, capillary rise may draw  
67 water from the groundwater aquifers to supply ET demands, declining water table. This two-way  
68 exchange between unsaturated soils and groundwater aquifers is important for the water table  
69 dynamics on regional scale.

70

71 [4] Furthermore, groundwater exchange with prairie pothole wetlands are complicated and critical  
72 in the PPR. Numerous wetlands known as potholes or sloughs provide important ecosystem  
73 services, such as providing wildlife habitats and groundwater recharge (Johnson et al., 2010).  
74 Shallow groundwater aquifers may receive water from or lose water to prairie wetlands depending  
75 on the hydrological setting. Depression-focused recharge generated by runoff from upland to  
76 depression contributes to sufficient amount of water input to shallow groundwater (5-40 mm/year).  
77 On the other hand, groundwater lateral flow exchange center of a wetland pond to its moist margin  
78 is also an important components in the wetland water balance (van der Kamp and Hayashi, 2009;



79 Brannen, et al., 2015; Hayashi et al., 2016). However, this groundwater-wetland exchange  
80 typically occurs on local scale (from 10 to 100 m) and thus, challenging to in current land surface  
81 models or climate models (resolution from 1 km to 100 km). Therefore, in this paper, we tend to  
82 focus on the groundwater dynamics in PPR on regional scale two-way exchange between aquifer  
83 and soils, rather than local scale groundwater-wetland exchange.

84

85 [5] Previous studies have suggest that substantial changes to groundwater interactions with above  
86 soils are likely to occur under climate change (Tremblay et al., 2011; Green et al., 2011; Ireson et  
87 al., 2013, 2015). Existing modeling studies on the impacts of climate change on groundwater are  
88 either at global or basin/location-specific scales (Meixner et al., 2016). Global-level groundwater  
89 studies focus on potential future recharge trends (Doll and Fiedler, 2008; Doll, 2009), yet coarse  
90 resolution analysis from global climate models (GCMs) provided little specificity to inform  
91 decision making. Basin-scale groundwater studies connect the climate with groundwater-flow  
92 models to understand the climate impacts on specific systems (Maxwell and Kollet, 2008; Kurylyk  
93 and MacQuarrie, 2013; Dumanski et al., 2015). However, a knowledge gap exists in predicting the  
94 effect of climate change over large regions (major river basins, states or group of states) (Green et  
95 al., 2011). The lack of climate-groundwater studies at regional scale may be due to two reasons:  
96 first, it is challenging to represent the two-way water exchange in coupled land surface and  
97 groundwater model at a regional scale; and second, a regional coupled land-hydrology model  
98 requires fine-resolution and good quality meteorological forcing, which needs to be further  
99 downscaled from GCMs.

100



101 [6] Recently, the two-way exchange has been implemented in coupled land surface – groundwater  
102 models (LSM-GW), but is still challenging to apply to cold regions. For example, Maxwell and  
103 Miller (2005) used a groundwater model (ParFlow) coupled with the Common Land Model (CLM).  
104 They found that the coupled and uncoupled model is very similar in simulated sensible heat flux  
105 (SH), ET, and shallow soil moisture (SM), but are different greatly in runoff and deep SM. This is  
106 perhaps because only downwards flow from soil to groundwater is considered. Later on, Kollet  
107 and Maxwell (2008) incorporated the ET effect on redistributing moisture upward from shallow  
108 water table depth (WTD) and found the surface energy partition is highly sensitive to a WTD  
109 ranging from 1 – 5 m. More recently, Niu et al. (2011) implemented a simple groundwater model  
110 (SIMGM, Niu et al., 2007), into the community Noah LSM with multi-parameterization options  
111 (Noah-MP LSM), by adding an unconfined aquifer at the bottom of soil layers. In order to  
112 reasonably capture the groundwater regime in the PPR under semi-arid and seasonally frozen soil  
113 climatic conditions, a coupled LSM-GW with the ability of addressing the two-way exchange  
114 between soil and groundwater as well as representing the freeze-thaw process is necessary.

115

116 [7] In addition to representing the two-way water exchange, the spatial heterogeneity of soil  
117 moisture and WTD requires meteorological forcing in high spatial resolution, that are not available  
118 from direct output of coarse resolution GCMs. Furthermore, great uncertainties of simulated  
119 precipitation stem from choice of convection parameterization schemes in GCMs (Sherwood et al.,  
120 2014; Prein et al., 2015). For example, convective precipitation, an important source of  
121 precipitation in the PPR in summer, its frequency, diurnal cycle and propagation are poorly  
122 captured by GCM's parameterizations (Rasmussen et al., 2017). An important approach to  
123 improve precipitation simulation is to use the convection-permitting model (CPM) (Ban et al.,



124 2014; Prein et al., 2015; Liu et al., 2017). The CPM uses a high spatial resolution in the atmosphere  
125 (usually under 5-km) to explicitly resolve convection and not activate convection parameterization  
126 schemes. On the other hand, CPMs can also improve the representation of fine-scale topography  
127 and spatial variations of surface fields (Prein et al., 2013). These CPM added-values provide an  
128 excellent opportunity to investigate groundwater evolution in the PPR.

129

130 [8] Therefore, the purpose of this paper is to investigate the hydrological changes in groundwater  
131 in PPR under climate change and understand the drivers for different hydrological processes. Our  
132 goal is to 1) simulate the two-way water exchange in the PPR using a coupled land-groundwater  
133 model, 2) capture changes in the groundwater regime under climate change, and 3) identify major  
134 climatic and land surface processes that contribute to these changes in the PPR. We use a  
135 deterministic distributed physical process-based LSM (Noah-MP LSM) coupled with a  
136 groundwater dynamics model, called the MMF model (developed by Fan et al. (2007) and Miguez-  
137 Macho et al. (2007)). The coupled Noah-MP-MMF model is driven by two sets of meteorological  
138 forcing for 13 years under current and future climate scenarios. These two sets of meteorological  
139 dataset are from a CPM dynamical downscale project using the Weather Research & Forecast  
140 (WRF) model with 4-km grid spacing in the Contiguous U.S. (WRF CONUS, Liu et al., 2017).

141

142 [9] The paper is structured as follow: Section 2 introduces the observational data for WTD in the  
143 PPR, the coupled Noah-MP-MMF model, and the meteorological forcing from WRF CONUS  
144 project. Section 3 evaluates the model simulated WTD timeseries against observations and shows  
145 the groundwater budget and hydrological changes due to climate change. Section 4 and 5 offer a  
146 broad discussion and conclusion to the paper.



147 2. Data and Methods

148 2.1 Observation data

149 [10] Groundwater observation data were obtained through several agencies: (1) the United States  
150 Geological Survey (USGS) National Water Information System in the U.S.  
151 ([https://waterdata.usgs.gov/nwis/uv?referred\\_module=gw&search\\_criteria=search\\_site\\_no&search](https://waterdata.usgs.gov/nwis/uv?referred_module=gw&search_criteria=search_site_no&search_criteria=site_tp_cd&submitted_form=introduction)  
152 [ch\\_criteria=site\\_tp\\_cd&submitted\\_form=introduction](http://aep.alberta.ca/water/programs-and-services/groundwater/groundwater-observation-well-network/default.aspx)), (2) the Alberta Environment  
153 ([http://aep.alberta.ca/water/programs-and-services/groundwater/groundwater-observation-well-](http://aep.alberta.ca/water/programs-and-services/groundwater/groundwater-observation-well-network/default.aspx)  
154 [network/default.aspx](https://www.wsask.ca/Water-Info/Ground-Water/Observation-Wells/)), (3) the Saskatchewan Water Security Agency  
155 (<https://www.wsask.ca/Water-Info/Ground-Water/Observation-Wells/>), and (4) the Manitoba  
156 Water Stewardship Division  
157 ([https://manitoba.maps.arcgis.com/apps/webappviewer/index.html?id=28fc1a53c9f8435d897501](https://manitoba.maps.arcgis.com/apps/webappviewer/index.html?id=28fc1a53c9f8435d897501724766a992)  
158 [724766a992](https://manitoba.maps.arcgis.com/apps/webappviewer/index.html?id=28fc1a53c9f8435d897501724766a992)). Initially, groundwater data from 166 wells were acquired, 40 in US, 56 from Alberta,  
159 61 from Saskatchewan, and nine from Manitoba.

160

161 [11] Despite the data acquired from a large number of wells, not all of the wells were suitable for  
162 the study of shallow groundwater and climate change. We used the following criteria to select  
163 qualified stations for our study and evaluate our model performance against these observations:

- 164 1) a sufficiently long record of groundwater measurement record during the simulation period;
- 165 2) minimal anthropogenic effects (such as pumping or irrigation);
- 166 3) unconfined aquifers with shallow groundwater levels (top 7 meters below surface).

167





168 These criteria reduced the observation data to the record of 11 well records, with one in Alberta,  
169 six in Saskatchewan and four in Minnesota (U.S.). **Table 1** summarizes the information for each  
170 selected well, and **Fig. 1(a)** shows the location of the wells in our study area.

171

172 **Fig. 1** (a) Topography of the Prairie Pothole Region (PPR) and station location of rain gauges (black dots) and  
173 groundwater wells (red diamonds); (b) Topography of the WRF CONUS domain, with the black box indicating the  
174 PPR domain.

175

176 **Table 1** Summary of the locations and aquifer type and soil type of the 11 selected wells.

177



178 2.2 Groundwater Scheme in Noah-MP

179 [12] In the present study, we used the community Noah-MP LSM (Niu et al. 2011; Yang et al.  
 180 2011), coupled with a GW model – the MMF model, (Fan et al. 2007; Miguez-Macho et al., 2007).  
 181 This coupled model has been applied in many regional hydrology studies in offline mode (Miguez-  
 182 Macho and Fan 2012; Martinez et al., 2016) as well as coupled with regional climate models  
 183 (Anyah et al., 2008; Barlage et al., 2015). We present here a brief introduction to the MMF  
 184 groundwater scheme, further details can be found in previous studies (Fan et al., 2007 and Miguez-  
 185 Macho et al., 2007).

186

187 [13] Fig. 2 is a diagram of the structure of Noah-MP soil layers and the MMF aquifer. The active  
 188 2-m soil in Noah-MP LSM consists of 4 layers whose thicknesses ( $dz$ ) are 0.1, 0.3, 0.6 and 1.0 m.  
 189 The MMF scheme defines explicitly an unconfined aquifer below the 2-m soil and an auxiliary  
 190 soil layer stretching to the WTD, which varies in space and time [m]. The thickness of this auxiliary  
 191 layer ( $z_{aux}$  [m]) is also variable, depending on the WTD:

$$192 \quad z_{aux} = \begin{cases} 1, & WTD \geq -3 \\ -2 - WTD, & WTD < -3 \end{cases} \quad (1)$$

193

194 [14] Fig. 2 shows the soil layers of NoahMP coupled MMP groundwater scheme. The vertical  
 195 fluxes include gravity drainage and capillary flux, solved from the Richards' equation,

$$196 \quad q = K_{SMC} * \left( \frac{\partial \Psi}{\partial z} - 1 \right), \quad K_{SMC} = K_{SAT} * \left( \frac{SMC}{SMCSAT} \right)^{2b+3}, \quad \Psi = \Psi_{SAT} * \left( \frac{SMCSAT}{SMC} \right)^b \quad (2)$$

197 where  $q$  is water flux between two adjacent layers [m/s],  $K_{SMC}$  is the hydraulic conductivity [m/s]  
 198 at certain soil moisture content  $SMC$  [m<sup>3</sup>/m<sup>3</sup>],  $\Psi$  is the soil capillary potential [m] and  $b$  is soil pore  
 199 size index. The subscript  $SAT$  denote saturated state whose values are from the default Noah-MP



200 soil tables. Therefore, the recharge flux from/to the layer above WTD,  $R$ , can be obtained  
201 according to WTD:

$$202 \quad R = \begin{cases} K_k * \left( \frac{\Psi_i - \Psi_k}{z_{soil(i)} - z_{soil(k)}} - 1 \right), & WTD \geq -2 \\ K_{aux} * \left( \frac{\Psi_4 - \Psi_{aux}}{(-2) - (-3)} - 1 \right), & -2 > WTD \geq -3 \\ K_{SAT} * \left( \frac{\Psi_{aux} - \Psi_{SAT}}{(-2) - (WTD)} - 1 \right), & WTD < -3 \end{cases} \quad (3)$$

203 In the first case, WTD is in the resolved soil layers and  $z_{soil}$  is the depth of soil layer with the  
204 subscript  $k$  indicating the layer containing WTD while  $i$  the layer above. The calculated water table  
205 recharge is then passed to the MMF groundwater routine.

206

207 [15] The change of groundwater storage in the unconfined aquifer considers three components:  
208 recharge flux, river flux, and lateral flows:

$$209 \quad \Delta S_g = (R - Q_r + \sum Q_{lat}) \quad (4)$$

210 where  $S_g$  [mm] is groundwater storage,  $Q_r$  [mm] is the water flux of groundwater-river exchange,  
211 and  $\sum Q_{lat}$  [mm] are groundwater lateral flows to/from all surrounding grid cells (Fan et al., 2007;  
212 Miguez-Macho et al., 2007). The groundwater lateral flow ( $\sum Q_{lat}$ ) is the total horizontal flows  
213 between each grid cell and its neighbouring grid cells, calculated from Darcy's law with the  
214 Dupuit–Forchheimer approximation (Fan and Miguez-Macho 2010), as:

$$215 \quad Q_{lat} = wT \left( \frac{h - h_n}{l} \right) \quad (5)$$

216 where  $w$  is the width of cell interface [m],  $T$  is the transmissivity of groundwater flow [m<sup>2</sup>/s],  $h$   
217 and  $h_n$  are the water table head [m] of local and neighboring cell, and  $l$  is the length [m] between  
218 cells.  $T$  depends on hydraulic conductivity  $K$  and WTD:



$$219 \quad T = \begin{cases} \int_{-\infty}^h K dz & WTD \geq -2 \\ \int_{-\infty}^{(z_{surf}-2)} K dz + \sum K_i * dz_i & WTD < -2 \end{cases} \quad (7)$$

220 For  $WTD < -2$ ,  $K$  is assumed to decay exponentially with depth,  $K = K_4 \exp(-z/f)$ ,  $K_4$  is the  
 221 hydraulic conductivity in the 4-th soil layer and  $f$  is the e-folding length and depends on terrain  
 222 slope. For  $WTD \geq -2$ ,  $i$  represents the number of layers between the water table and the 2-m bottom  
 223 and  $z_{surf}$  is the surface elevation.

224

225 [16] The river flux ( $Q_r$ ) is also represented by a Darcy's law-type equation, which is the gradient  
 226 between the groundwater head, local riverbed depth and parameterized river conductance:

$$227 \quad Q_r = RC \cdot (h - z_{river}) \quad (8)$$

228 with  $z_{river}$  is the depth of river bed [m] and  $RC$  is dimensionless river conductance, which depends  
 229 on the slope of the terrain and equilibrium water table ( $eqzwt$ , [m]). Eq. (7) is a simplification  
 230 which uses  $z_{river}$  rather than the water level in the river and, for this study, we only consider one-  
 231 way discharge from groundwater to rivers. Finally, the change of WTD is calculated as the total  
 232 fluxes fill or drain the pore space between saturation and the equilibrium soil moisture state  
 233 ( $SMCEQ$  [ $m^3/m^3$ ]) in the layer containing WTD:

$$234 \quad \Delta WTD = \frac{\Delta S_g}{(SMCSAT - SMCEQ)} \quad (9)$$

235 If  $\Delta S_g$  is greater than the pore space in the current layer, the soil moisture content of current layer  
 236 is saturated and the WTD rises to the layer above, updating the soil moisture content in the layer  
 237 above as well. Vice versa for negative  $\Delta S_g$  as water table declines and soil moisture decreases.

238



239 **Fig. 2** Structure of the Noah-MP LSM coupled with MMF groundwater scheme, the top 2-m soil is consist of 4 layers  
240 whose depth are 0.1, 0.3, 0.6 and 1.0 m. An unconfined aquifer is added below the 2-m boundary, including an  
241 auxiliary layer and the saturated aquifer. Positive flux of  $R$  denotes downward transport. Two water table are shown,  
242 one within the 2-m soil and one below, indicating that the model is capable to deal with shallow as well as deep water  
243 table.  
244



245 2.3 Forcing Data

246 [17] We use the atmospheric forcing from the WRF CONUS simulation (Liu et al. 2017) to drive  
247 the Noah-MP-MMF model. The WRF CONUS simulation consists of two parts: one is the current  
248 climate scenario from Oct 2000 to Sep 2013, downscaled from ERA-Interim reanalysis data; the  
249 other is the pseudo global warming (PGW) as the future climate scenario, which adds a delta  
250 climate change signal derived from an ensemble of CMIP5 models.

251

252 [18] **Fig. 3** shows the annual precipitation in the PPR from 4-km WRF CONUS from the current  
253 climate and 32-km North America Regional Reanalysis (NARR, another reanalysis dataset  
254 commonly used for land surface model forcing). Both datasets show similar annual precipitation  
255 pattern and bias patterns compared to observations: underestimating of precipitation in the east  
256 and overestimating in the west. However, the WRF CONUS shows significant improvement of  
257 percentage bias in precipitation ( $((\text{Model}-\text{Observation})/\text{Observation})$ ) over the western PPR. To also  
258 address the consistency of using the same source of data for current and future climate, we believe  
259 the 4-km WRF CONUS is the best available dataset to drive the coupled land-groundwater model.

260

261 **Fig. 3** Evaluation of the annual precipitation from WRF CONUS (top) and NARR (bottom) against rain gauge  
262 observation.

263

264 [19] For the future climate study, the precipitation and temperature of the PGW climate forcing  
265 are shown in **Fig. 4** and **Fig. 5**. The WRF CONUS projects more precipitation in the PPR, except  
266 in the southeast of the domain in summer, where it shows a precipitation reduction of about 50 to  
267 100 mm. On the other hand, the WRF CONUS projects strongest warming occurring in the  
268 northeast PPR in winter (**Fig. 5**) – warming of about 6–8 °C. Another significant warming signal



269 occurs in summer in the southeast of domain, corresponding to the reduction of future precipitation,

270 as seen in **Fig. 4**.

271

272 **Fig. 4** Seasonal accumulated precipitation from current climate scenario(CTRL), future climate scenario (PGW) and

273 projected change (PGW-CTRL) in the forcing data.

274

275 **Fig. 5** Same as **Fig. 4** but for 2-m air temperature.

276



277 2.4 Model Setup

278 [20] The two Noah-MP-MMF simulations representing the current climate and future climate are  
279 denoted as CTRL and PGW, respectively. The CTRL simulation starts from Oct 2000 and end in  
280 Sep 2013, consist of 13 water years. The PGW simulation corresponds to the same period of time,  
281 but under climate change scenario at the end of 21<sup>st</sup> century.

282

283 [21] The model's initial groundwater is from a global 1-km equilibrium groundwater map (Fan et  
284 al., 2013) and the equilibrium soil moisture for each soil layer is calculated at the first model  
285 timestep with climatology recharge, spinning up for 500 years. Since the model domain is at a  
286 different resolution than the input data, the appropriate initial WTD at 4-km may be different than  
287 the average at 1-km. To properly initialize the simulation, we spin up the model using the forcing  
288 of current climate (CTRL) for the years from 2000 to 2001 repeatedly (in total 4 loops).

289

290 [22] Due to different data sources, the default soil types along the boundary between the U.S. and  
291 Canada are discontinuous. Thus, we use the global 1-km fine soil data (Shangguan et al., 2014,  
292 <http://globalchange.bnu.edu.cn/research/soilw>) in our study region. The coupled Noah-MP-MMF  
293 groundwater model is configured using the default 2-m depth with 4 layers. The soil properties for  
294 the aquifer use the same properties as the lowest soil layer.





## 295 3. Results

## 296 3.1 Comparison with groundwater observations

297 [23] According to the locations of 11 groundwater wells in **Table 1**, the simulated WTD from the  
298 closest model grid points are extracted. **Fig. 6** shows the observed WTD (black plus) and simulated  
299 monthly WTD (blue lines) in the study domain. The model successfully captures the annual cycle  
300 of WTD, which rises in spring and early summer, because of snowmelt and rainfall recharge, and  
301 declines in summer and fall, because of high ET, and in winter because of frozen near-surface soil.  
302 In all observations, the timing of water table rising and dropping is well simulated, as the timing  
303 and amount of infiltration and recharge in spring is controlled by the freeze-thaw processes in  
304 seasonally frozen soil. These processes are reasonably captured by the frozen soil scheme in Noah-  
305 MP LSM (Niu et al., 2006; Niu et al., 2011).

306

307 **Fig. 6** WTD (m) from 11 groundwater wells and the model simulation results in the PPR.

308

309 [24] On the other hand, the model simulated WTD seasonal variation is smaller than observations.  
310 The small seasonal variation could be due to the mismatch between the lithology from the  
311 observational surveys and the soil types in the model grids. As mentioned in Section 2.2, the soil  
312 properties of the unconfined aquifer are the same as the bottom layer of the resolved 2-m soil  
313 layers. While sand and gravel are the dominant lithology in most of the sites, except for silt in  
314 Crater Lake, they are mostly clay and loam in the model (Table 1). For sandy soil reported in most  
315 of the sites, fast responses to infiltration and drainage lead to large water table fluctuations,  
316 whereas, in the model, clay and loam soil allows low permeability and smoothens responses to  
317 recharge and capillary effects. This shortcoming of the model was also reported in a study taken  
318 place in the Amazon rainforest (Miguez-Macho et al., 2012).



319

320 [25] Despite mismatches between the model and sites in topography and soil properties at high  
321 spatial resolution, this out-of-the-box simulation of the Noah-MP MMF groundwater scheme  
322 shows reasonable results, as shown in **Fig. 6**. Therefore, we consider the simulated WTD  
323 satisfactory in the mean, seasonal, and interannual dynamics, and are reliable for further study of  
324 climate change impacts on groundwater in the PPR.

325



## 326 3.2 Climate change signal in Groundwater fluxes

327 [26] The MMF groundwater model simulates three components in the groundwater water budget,  
328 the recharge flux ( $R$ ), lateral flow ( $Q_{lat}$ ), and discharge flux to rivers ( $Q_r$ ). Because the topography  
329 is usually flat in the PPR, the magnitude of groundwater lateral transport is very small ( $Q_{lat}$  less  
330 than 5 mm per year). On the other hand, the shallow water table in the PPR region is higher than  
331 the local river bed, thus, the  $Q_r$  term is always negative and discharging from groundwater aquifers  
332 to rivers. As a result, the recharge term is the major contributor to the groundwater storage in the  
333 PPR, and its variation (usually between -100 to 100 mm) dominates the timing and amplitude of  
334 the water table dynamics. The seasonal accumulated total groundwater fluxes in the PPR  
335 ( $R+Q_{lat}+Q_r$ ) are shown in **Fig. 7**. The positive flux (blue) means the groundwater aquifer is  
336 gaining water, causing the water table to rises; and the negative flux (red) indicates the aquifer is  
337 losing water and the water table is declining.

338

339 **Fig. 7** Same as **Fig. 3** but for total groundwater fluxes ( $R+Q_{lat}-Q_r$ ).

340

341 [27] Under current climate, the total groundwater flux show strong seasonal fluctuations,  
342 consistent with the WTD timeseries shown in **Fig. 6**. On average, in fall (SON) and winter (DJF),  
343 there is a 20-mm negative recharge, driven by the capillary effects and drawing water from aquifer  
344 to dry soil above. Spring (MAM) is usually the season with a strong positive recharge because  
345 snowmelt provides a significant amount of water, and soils thawing allow infiltration. The large  
346 amount of snowmelt water contributes to more than 100 mm of positive recharge in the eastern  
347 domain. It is until summer (JJA), when strong ET depletes soil moisture and results in about 50  
348 mm of negative recharge.

349



350 [28] Under future climate, the increased PR in fall and winter leads to wetter upper soil layers,  
351 resulting in a net positive recharge flux (PGW – CTRL in SON and DJF). However, the PGW  
352 summer is impacted by increased ET under a warmer and drier climate, due to higher temperature  
353 and less PR. As a result, the groundwater uptake by the capillary effect is more critical in the future  
354 summer. Furthermore, there is a strong east-to-west difference in the total groundwater flux change  
355 from PGW to CTRL. In the eastern PPR, the change in total groundwater flux exhibits obvious  
356 seasonality while the model projects persistent positive groundwater fluxes in the western PPR.  
357



## 358 3.3 Water budget analysis

359 [29] **Fig. 8** and **Fig. 9** show the water budget analysis for the eastern and western PPR (divided by  
360 the dotted line in 103° W in Fig. 7), respectively. Four components are presented in the figures,  
361 i.e. (1) PR and ET; (2) surface and underground runoff (*SFCRUN* and *UDGRUN*); and surface  
362 snowpack; (3) the change of soil moisture storage and (4) groundwater fluxes and the change of  
363 storage. In current and future climate, these budget terms are plotted in annual accumulation ((a)  
364 and (b) for CTRL and PGW), whereas their difference are plotted in each month individually ((c)  
365 for PGW-CTRL).

366

367 [30] During snowmelt infiltration and rainfall events when ET demand is low, water infiltrates into  
368 the top soil layer, travels through the soil column and exits the bottom of the 2-m boundary, hence,  
369 the water table rises. During the summer dry season, ET is higher than PR and the soil layers lose  
370 water through ET, therefore, the capillary effect takes water from the underlying aquifer and the  
371 water table declines. In winter, the near-surface soil in the PPR is seasonally frozen, thus, a  
372 redistribution of subsurface water to the freezing front results in negative  $Q_{drain}$ , and the water  
373 table declines.

374

375 [31] In the eastern PPR, the effective precipitation (PR-ET) is found to increase from fall to spring,  
376 but decrease in summer in PGW (**Fig. 8(1c)**). Warmer falls and winters in PGW, together with  
377 increased PR, not only results in later time for snow accumulation and earlier for melting, but also  
378 changes the precipitation partition – more as rain and less as snow. This warming causes up to 20  
379 mm of snowpack loss (**Fig.8(2c)**). The underground runoff starts much earlier in PGW (December)  
380 (**Fig.8(2b)**) than in CTRL (February) (**Fig.8(2a)**). On the other hand, the warming in PGW also



381 changes the partitioning of soil ice and soil water in subsurface soil layers (**Fig. 8(3c)**). For late  
382 spring in PGW, the springtime recharge in the future is significantly reduced due to early melting  
383 and less snowpack remaining (**Fig. 8(4c)**). In the PGW summer, reduced PR (50 mm less) and  
384 higher temperatures (8 °C warmer) lead to reduction in total soil moisture, and a stronger negative  
385 recharge from the aquifer. Therefore, the increase of recharge from fall to early spring compensates  
386 the recharge reduction due to stronger ET in summer in the eastern PPR, and changes little in the  
387 annual mean groundwater storage (1.763 mm per year).

388

389 **Fig. 8** Water budget analysis in the eastern PPR in (a) CTRL, (b) PGW and (c) PGW – CTRL. Water budget terms  
390 include: (1) *PR & ET*, (2) surface snow, surface runoff and underground runoff (*SNOW*, *SFCRUN*, and *UDGRUN*),  
391 (3) change of soil moisture storage (soil water, soil ice and total soil moisture,  $\Delta SMC$ ) and (4) groundwater fluxes  
392 and the change of groundwater storage ( $R$ ,  $Q_{lat}$ ,  $Q_r$ ,  $\Delta S_g$ ). The annual mean soil moisture change (PGW-CTRL) is  
393 shown with black dashed line in (3). The Residual term is defined as  $Res = (R+Q_{lat}-Q_r)-\Delta S_g$  in (4). Note that in (a)  
394 and (b) the accumulated fluxes and change in storage are shown in lines, whereas in (c) the difference in (PGW-CTRL)  
395 is shown for each individual month in bars.

396

397 [32] These changes in water budget components in the western PPR (**Fig. 9**) are similar to those  
398 in the eastern PPR (**Fig. 8**), except in summer. The reduction in summer PR in the western the PPR  
399 (less than 5 mm reduction) is not as obvious as that in the eastern PPR (50 mm reduction) (**Fig. 4**).  
400 Thus, annual mean total soil moisture in future is about the same as in current climate (Fig. 9(3c))  
401 and results in little negative recharge in PGW summer (**Fig. 9(4c)**). Therefore, the increase in  
402 annual recharge is more significant (10 mm per year), an increase of about 50% of the annual  
403 recharge in the current climate (20 mm per year) (**Fig. 9(4c)**).

404

405 **Fig. 9** Same as **Fig. 8**, but for the western PPR.



406

407 [33] In both the eastern and western PPR, the water budget components for the groundwater aquifer  
408 are plotted in **Fig. 8(4)** and **Fig. 9 (4)**, with the changes of each flux (PGW-CTRL) printed at the  
409 bottom. The groundwater lateral flow is a small term in areal average and has little impact on the  
410 groundwater storage. Nearly half of the increased recharge in both the eastern and western PPR is  
411 discharged to river flux ( $Q_r = 2.26$  mm out of  $R = 4.15$  mm in the eastern PPR and  $Q_r = 5.20$  mm  
412 out of  $R = 10.72$  mm in western PPR). Therefore, the groundwater storage change in the eastern  
413 PPR (1.76 mm per year) is not as great as that in the western PPR (5.39 mm per year).

414

415 [34] These two regions of the PPR show differences in hydrological response to future climate  
416 because of the spatial variation of the summer PR. As shown in both **Fig. 4** (PGW-CTRL), **Fig.**  
417 **8(1)** and **Fig. 9(1)**, the reduction of future PR in summer in the eastern PPR is significant (50 mm).  
418 The spatial difference of precipitation changes in the PPR further results in the recharge increase  
419 doubling in the western PPR compared to the eastern PPR.

420



## 421 4. Discussion

## 422 4.1 Simulated WTD sensitivity to soil property parameters

423 [35] In Section 3.1, we show that model is capable of simulating the mean WTD, yet  
424 underestimates its seasonal variation due to mismatches between model default soil type and the  
425 soil properties in the observational wells. To test this theory, an additional simulation, REP, is  
426 conducted by replacing the default soil types in the 11 sites with the dominant soil types reported  
427 from observational surveys. The time series of the REP and default (MOD) are shown in Fig. 10  
428 and a summary of the mean and standard deviation of the two simulations are provided in Table 2.  
429

430 **Fig. 10** Same as Fig. 6, the time series of simulated WTD from both default model (MOD) and replacing soil type  
431 simulation (REP).

432

433 [36] The REP simulation with soil type from observational surveys show two sensitive signals: (1)  
434 REP WTD (red lines in Fig. 10) are shallower than the default simulation; (2) REP WTD timeseries  
435 shows stronger seasonal variation. These two signals can be explained by the WTD equation in  
436 the MMF scheme:

$$437 \quad \Delta WTD = \frac{\Delta(R + Q_{lat} + Q_r)}{(SMCSAT - SMCEQ)} \quad (10)$$

438 where  $SMCMAX$  is the maximum soil moisture capacity in the current layer, a parameter  
439 determined by the soil type. Eq. (10) represents that the change of WTD in a period of time is  
440 calculated by the total groundwater fluxes,  $\Delta(R + Q_{lat} + Q_r)$ , divided by the available soil  
441 moisture capacity of current layer ( $SMCMAX - SMCEQ$ ). In REP simulation, the  $SMCSAT$   
442 parameters for the dominant soil type in observational sites (sand/gravel) is smaller than those in  
443 default model grids (clay loam, sandy loam, loam, loamy sand, etc.). Therefore, changing the





444 *SMCSAT* is essentially altering the storage in the aquifer and soil in this model grid. Given the  
445 same amount of groundwater flux, in the REP simulation, the mean WTD is higher than the default  
446 run and the seasonal variation is stronger, hence, increasing the WTD seasonal variation.

447

448 [37] We show the REP simulation in order to prove our theory in Section 3.1, that the simulated  
449 WTD is sensitive to parameters associated soil properties, and thus could be improved by obtaining  
450 more realistic soil maps. But in the REP simulation, we replaced soil type with observations only  
451 at these 11 site locations because the geological survey data in high resolution and large area extent  
452 is not yet available for the whole PPR. Future development in fine-scale soil properties will  
453 hopefully improve the WTD simulation.

454

#### 455 4.2 Climate change Impacts on Groundwater Hydrologic Regime

456 [38] Climate change induced warming in high-latitudes winter and increased precipitation,  
457 including a higher liquid fraction, in PGW winter results in later snow accumulation, higher winter  
458 recharge and earlier melting in spring. Such changes in snowpack loss have been hypothesized in  
459 mountainous as well as high-latitude regions (Taylor et al 2013; Ireson et al., 2015; Meixner et al.,  
460 2016; Musselman et al., 2017). On an annual basis, the coupled model projects substantial  
461 increases in recharge, 25% in eastern and 50% in western PPR.

462

463 [39] In addition to the amount of recharge, the shift of recharge season is also noteworthy. Under  
464 current climate in spring, soil thawing (in March) is generally later than snowmelt (in February)  
465 by a month in the PPR. Thus, the snowmelt water in pre-thaw spring would either re-freeze after  
466 infiltrating into partially frozen soil or become surface runoff. Under the PGW climate, the warmer



467 winter and spring allows snowmelt and soil thaw to occur earlier in the middle of winter (in January  
468 and February, respectively). As a result, the recharge season starts earlier in December, and last  
469 longer until June, results in longer recharge season but with lower recharge rate.

470

471 [40] Future projected stronger evapotranspiration demand in summer desiccates upper soil layers,  
472 resulting in more water uptake from aquifers to subsidize upper dry soil in the future summer. This  
473 groundwater transport to upper soil layers is similar to the “buffer effect” documented in an offline  
474 study in the Amazon rainforest. In both Amazon rainforest and the PPR, shallow water tables exist  
475 in the critical zone from 1 to 5 meters below surface (Kollet and Maxwell, 2008; Fan , 2015) and  
476 could exert strong influence on land energy and moisture fluxes feedback to the atmosphere.  
477 Previous coupled atmosphere-land-groundwater studies at 30-km resolution showed that  
478 groundwater could support soil moisture during summer dry period, but has little impacts on  
479 precipitation in Central U.S. (Barlage et al., 2015). It would be an interesting topic to study the  
480 integrated impacts of shallow groundwater to regional climate in the convection permitting  
481 resolution (resolution < 5-km).

482



483 Conclusion

484 [41] The fluctuation of the shallow WTD and recharge are closely related to the semi-arid climatic  
485 condition and seasonally frozen soil in the Prairie Pothole Region (PPR). The freeze-thaw  
486 processes and effective precipitation are critical factors to the magnitude and timing for snowmelt  
487 infiltration in spring, groundwater uptake in summer and moisture redistribution in winter. The  
488 two-way water exchange between soil layers and groundwater aquifer is essential to the  
489 groundwater regime in the PPR.

490

491 [42] Previous works on modeling climate change impacts on groundwater, at both global or local  
492 basin scale, have been limited by precipitation uncertainties stemming from choice of convection  
493 parameterization in GCMs. (Green et al., 2011; Kurylyk et al., 2013; Taylor et al., 2013; Smerdon  
494 2017). Additionally, these models typically neglected the two-way exchange of water flux between  
495 soil and aquifers, which has been a historical simplification in the coupled land-groundwater model.

496

497 [43] In our study, a coupled land-groundwater model is applied to simulate the interaction between  
498 the groundwater aquifer and soil moisture in the PPR. The climate forcing is from a dynamical  
499 downscaling project (WRF CONUS), which uses the high spatial resolution convection-permitting  
500 model (CPM). To our knowledge, this is the first study applying convection-permitting regional  
501 climate model (RCM) forcing in a hydrology study. The goal of this study is to investigate the  
502 groundwater responses to climate change, and to identify the major processes that contribute to  
503 these responses in the PPR. We have three main findings:

504



505 [44] (1) the coupled land-groundwater model shows reliable simulation of mean WTD, however  
506 underestimates the seasonal variation of the water table against well observations. This is mostly  
507 due to the mismatches of soil types between groundwater sites and corresponding model grid  
508 points. This mismatch comes from inadequate information of aquifer parameters, which are the  
509 same as those for the lowest soil layer. We further demonstrated in an additional simulation (REP)  
510 by replacing the default soil type with observational “true” value that the simulated WTD is  
511 sensitive to soil type parameters, and the simulated WTDs were improved in both mean and  
512 seasonal variation. However, inadequacy of soil properties in deeper layer ( $< -2$  m) is still a  
513 limitation.

514

515 [45] (2) In general, recharge markedly increases due to projected increased PR, particularly from  
516 fall to spring under future climate condition. Strong east-west spatial variation exists in the annual  
517 recharge increases, 25% in the eastern and 50% in the western PPR. This is due to the significant  
518 projected PR reduction in PGW summer in the eastern PPR but little change in the western PPR.  
519 This PR reduction results in stronger ET, which draws more groundwater uptake due to the  
520 capillary effect, results in negative recharge in the summer. As a results, the increased recharge  
521 from fall to spring is consumed by ET in summer, and results in little change in groundwater  
522 storage in the eastern PPR, while higher storage in the western PPR.

523

524 [46] (3) The timing of infiltration and recharge are critically impacted by the changes in  
525 temperature-induced freeze-thaw processes. Increased precipitation, combined with higher winter  
526 temperatures, results in later snow accumulation/soil freezing, partitioned more as rain than snow,  
527 and earlier snowmelt/soil thaw. This leads to substantial loss of snowpack, shorter frozen soil



528 season, and higher permeability in soil allowing infiltration in the future winter. Additionally, late  
529 accumulation/freezing and early melting/thawing leads to an early start of a longer recharge season  
530 from December to June, but with a lower recharge rate.

531

532 [47] Our study has some limitations where future studies are needed:

533 (1) Despite the large number of groundwater wells in PPR, only a few are suitable for long-term  
534 evaluation, due to data quality, anthropogenic pumping, and length of data record. As remote  
535 sensing techniques advance, observing terrestrial water storage anomalies derived from the  
536 GRACE satellite may provide substantial information on WTD, although the GRACE information  
537 needs to be downscaled to a finer scale before comparisons can be made with regional hydrology  
538 models at km-scale (Pokhrel et al., 2013).

539

540 [48] (2) This study is an offline study of climate change impacts on groundwater. It is important  
541 to investigate how shallow groundwater in the earth's critical zone could interact with surface  
542 water and energy exchange to the atmosphere and affect regional climate. This investigation would  
543 be important to the central North America region (one of the land atmosphere coupling "hot spots",  
544 Koster et al., 2004 ).

545



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557



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Table and Figure

**Table 1.** Summary of the locations and aquifer type and soil type of the 11 selected wells.

WELL NAME	WELL ID	PROV	LAT	LON	ELEV	ELEV_MOD	AQUIFER TYPE	LITHOLOGY	SOIL_MOD
Narrow Lake	252	AB	54.600	-113.631	640.0	701.0	Surficial	Sand	Clay loam
Beauval	Beau	SK	55.117	-107.745	434.3	446.5	Intertill	Sand	Sandy loam
Crater lake	Crat	SK	50.953	-102.463	524.1	522.8	Intertill	Sand/Gravel/Silt	Loam
Duck lake	Duck	SK	52.921	-106.233	502.9	501.729	Surficial	Sand	Loamy sand
Forget	Forg	SK	49.705	-102.863	606.5	606.0	Surficial	Sand	Sandy loam
Simpson 14	SI14	SK	51.457	-105.193	496.6	493.3	Surficial	Sand	Sandy loam
Yorkton No.517	Y517	SK	51.173	-102.509	513.6	511.2	Surficial	Sand/Gravel	Loam
Prairie Island	PI98-14	MN	44.698	92.705	209.7	247.6	Surficial	Sand/Gravel	Loamy sand
CRN Well	WLN03	MN	45.983	95.203	410.7	411.4	Surficial	Sand/Gravel	Sandy loam
Glacial Ridge Well	G15	MN	47.635	96.254	351.7	344.2	Surficial	Sand/Gravel	Loam sand
Glacial Ridge Well	E03	MN	47.738	96.235	341.9	336.2	Surficial	Sand/Gravel	Sandy loam

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716 **Table 2.** Summary of mean and standard deviation (in the parenthesis) of WTD from 11  
717 groundwater wells, from observation records, default model (MOD) and replacing “true” soil  
718 properties simulation (REP). Bold texts show improvement in the mean and seasonal variation of  
719 WTD.  
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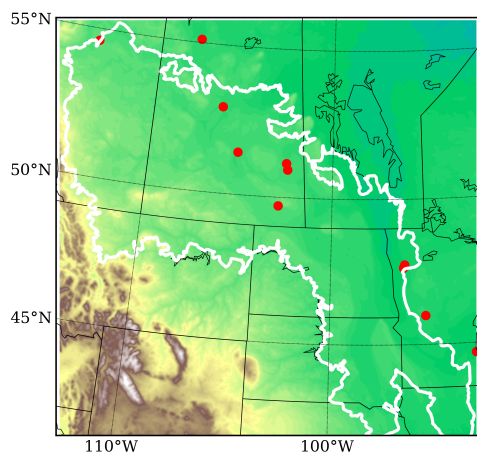
STATION	BEAU	NARR 252	DUCK	SI14	FORG	Y517	CRAT	GWR E03	GRW G15	CRN WLN03	PI98- 14
OBS	-3.78 (0.44)	-2.31 (0.28)	-3.66 (0.54)	-2.03 (0.34)	-2.28 (0.33)	-2.87 (0.80)	-4.33 (1.10)	-2.44 (0.39)	-1.94 (0.39)	-2.04 (0.24)	-4.31 (0.53)
MOD	-4.85 (0.56)	-4.81 (0.60)	-4.21 (0.42)	-2.61 (0.18)	-2.70 (0.19)	-3.97 (0.46)	-3.97 (0.40)	-2.29 (0.21)	-2.32 (0.20)	-2.18 (0.18)	-4.52 (0.12)
REP	<b>-4.20</b> <b>(0.32)</b>	<b>-3.75</b> <b>(0.51)</b>	<b>-3.54</b> (0.61)	<b>-1.82</b> (0.27)	<b>-1.66</b> <b>(0.46)</b>	<b>1.98</b> (0.32)	<b>-3.64</b> (0.28)	<b>-1.96</b> <b>(0.40)</b>	<b>-2.26</b> <b>(0.22)</b>	<b>-1.88</b> <b>(0.43)</b>	<b>-4.43</b> <b>(0.16)</b>

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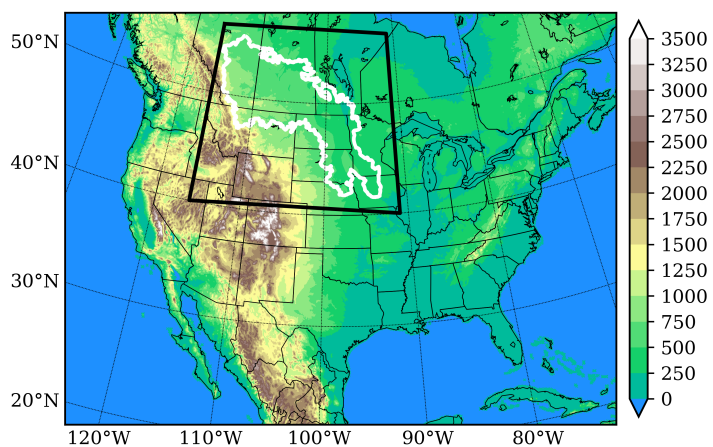


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724 (a)



(b)



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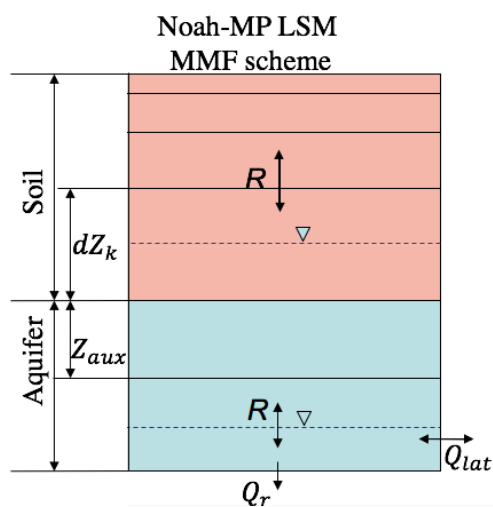
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**Fig. 1** (a) Topography of the Prairie Pothole Region (PPR; white outline) and station location of rain gauges (gray dots) and groundwater wells (red dots); (b) Topography of the WRF CONUS domain, the black box indicates the PPR domain.



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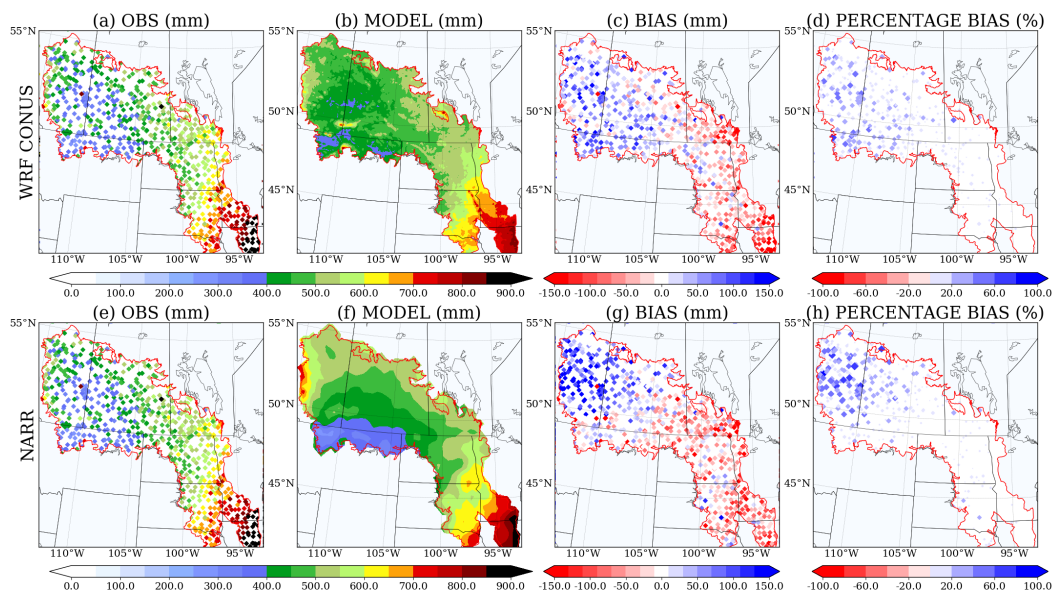
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731 **Fig. 2** Structure of the Noah-MP LSM coupled with MMF groundwater scheme, the top 2-m soil is consist of 4 layers  
732 whose depth are 0.1, 0.3, 0.6 and 1.0 m. An unconfined aquifer is added below the 2-m boundary, including an  
733 auxiliary layer and the saturated aquifer. Positive flux of  $R$  denotes downward transport. Two water table are shown,  
734 one within the 2-m soil and one below, indicating that the model is capable to deal with shallow as well as deep water  
735 table.

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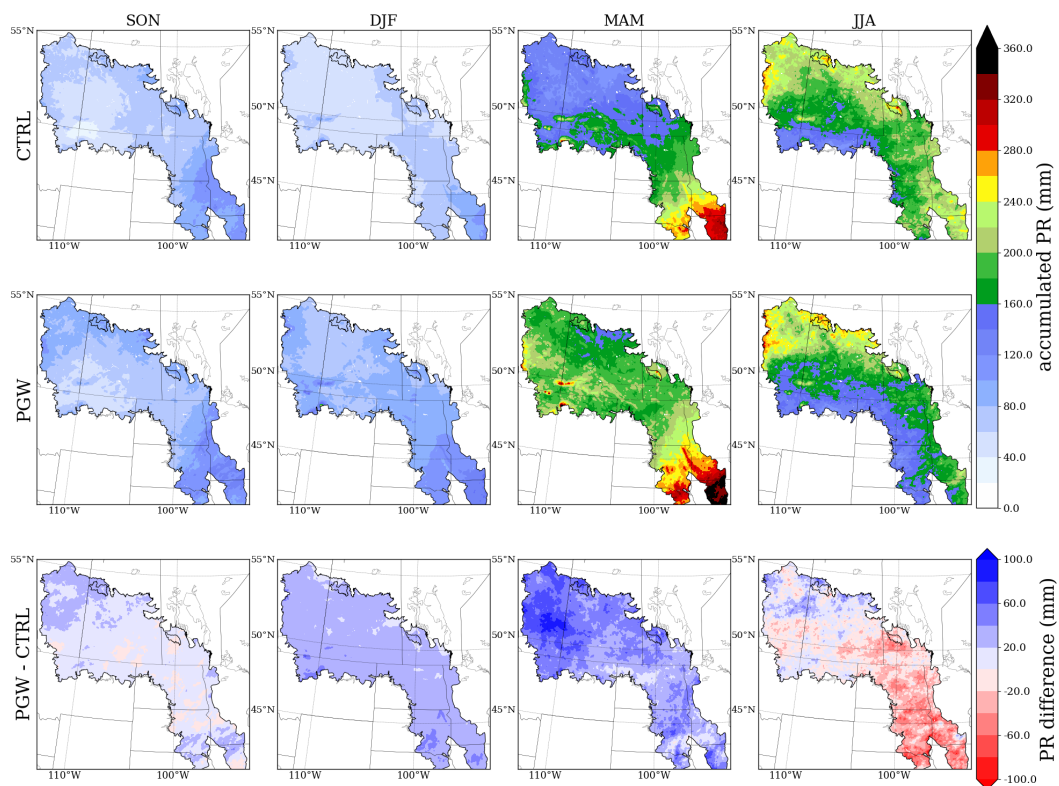
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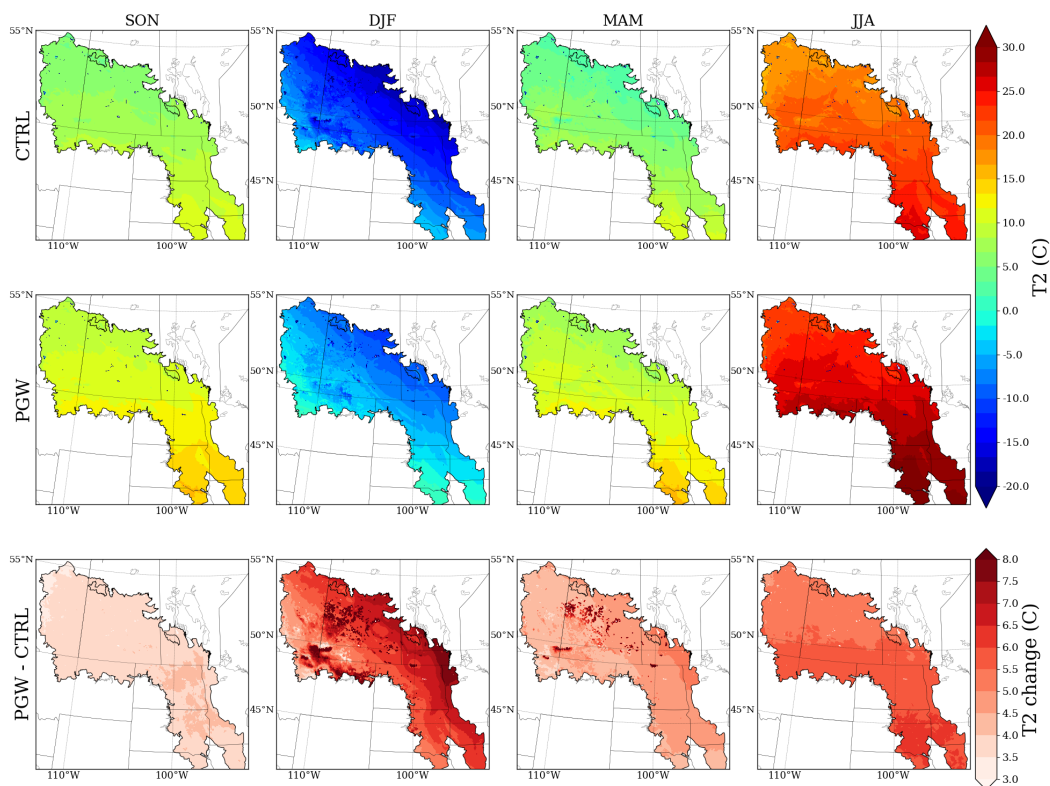
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**Fig. 3** Evaluation of the annual precipitation from two model products (b, f), WRF CONUS and NARR against rain gauge observation (a, e), their bias (c, g) and percentage bias (d, h).



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744 **Fig. 4** Seasonal Accumulated precipitation from current climate (CTRL, top), future climate (PGW, middle) and  
745 projected change (PGW-CTRL, bottom) in forcing data.  
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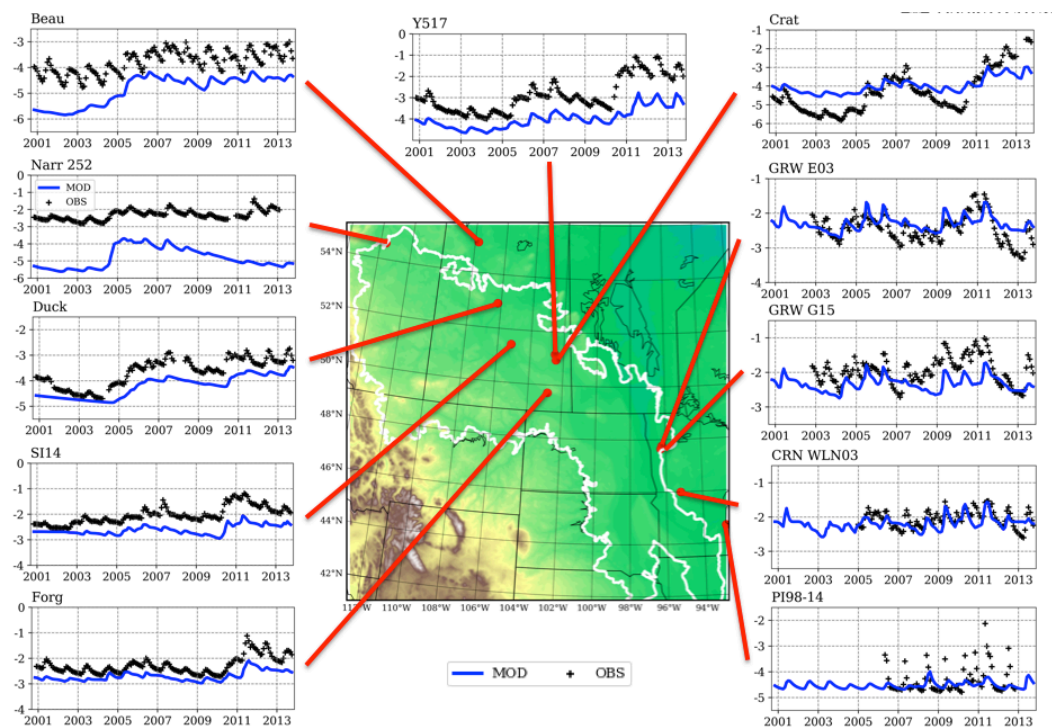


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**Fig. 5** Seasonal temperatures from current climate (CTRL, top), future climate (PGW, middle) and projected change (PGW-CTRL, bottom) in forcing data.

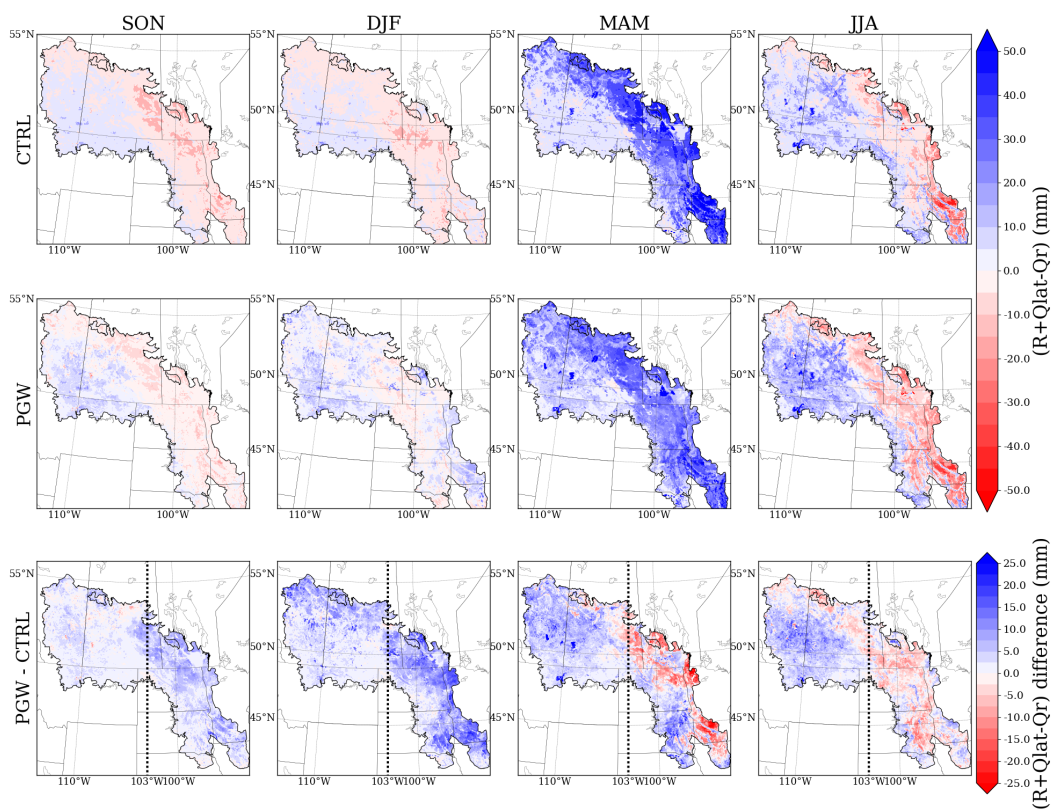


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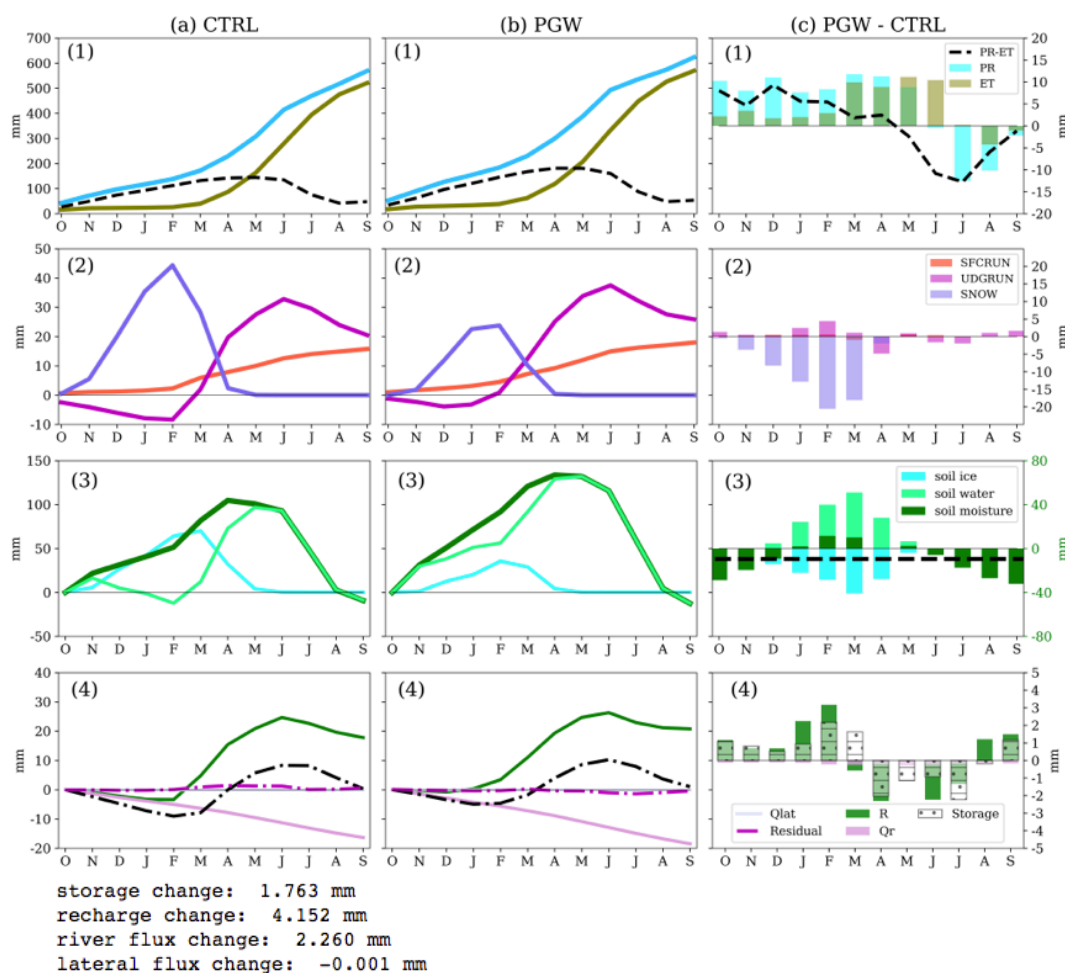
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**Fig. 6** WTD (m) from 11 groundwater wells and the model simulation results in PPR.



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**Fig. 7** Seasonal accumulated total groundwater fluxes ( $R+Q_{lat}+Q_r$ ) for current climate (CTRL, top), future climate (PGW, middle) and projected change (PGW-CTRL, bottom) in forcing data. Black dashed lines in PGW-CTRL separate the PPR into eastern and western halves.



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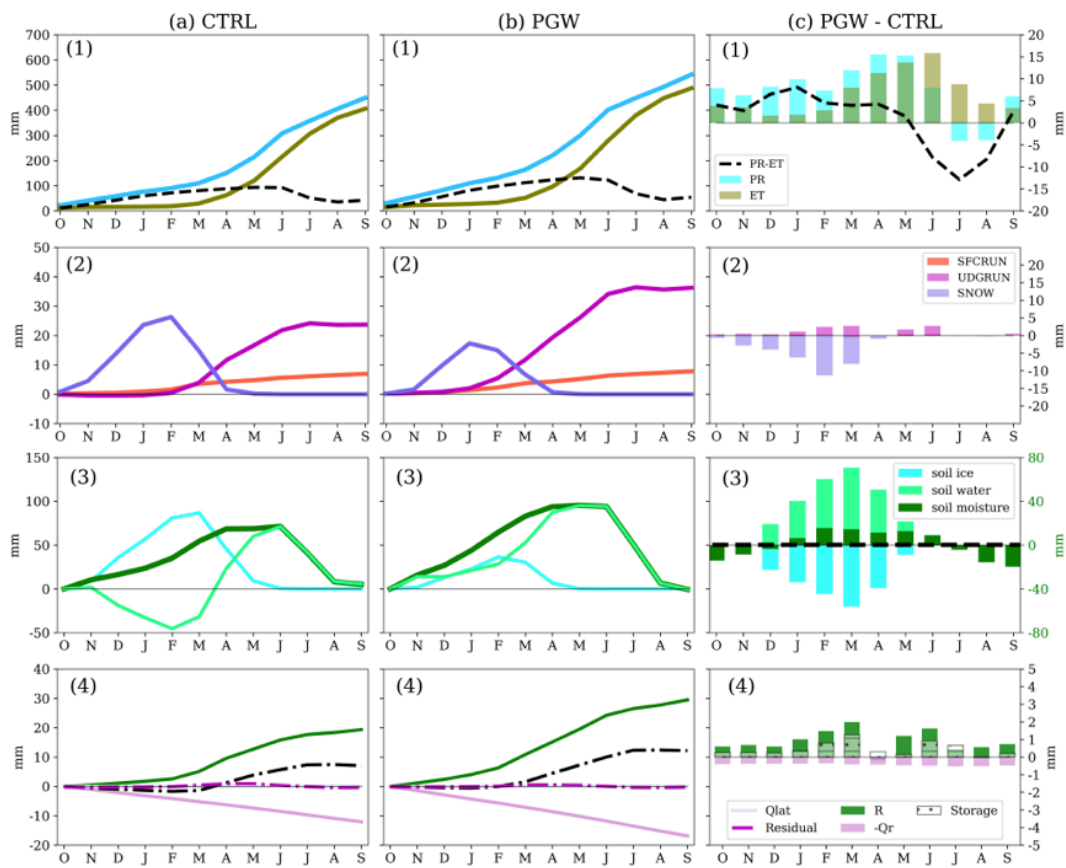
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**Fig. 8** Water budget analysis in the eastern PPR in (a) CTRL, (b) PGW and (c) PGW – CTRL. Water budget terms include: (1) *PR* & *ET*, (2) surface snow, surface runoff and underground runoff (*SNOW*, *SFCRUN*, and *UDGRUN*), (3) change of soil moisture storage (soil water, soil ice and total soil moisture,  $\Delta SMC$ ) and (4) groundwater fluxes and the change of groundwater storage ( $R$ ,  $Q_{lat}$ ,  $Q_r$ ,  $\Delta S_g$ ). The annual mean soil moisture change (PGW-CTRL) is shown with black dashed line in (3). The Residual term is defined as  $Res = (R + Q_{lat} - Q_r) - \Delta S_g$  in (4). Note that in (a) and (b) the accumulated fluxes and change in storage are shown in lines, whereas in (c) the difference in (PGW-CTRL) is shown for each individual month in bars.



storage change: 5.390 mm  
 recharge change: 10.727 mm  
 river flux change: 5.207 mm  
 lateral flux change: 0.000 mm

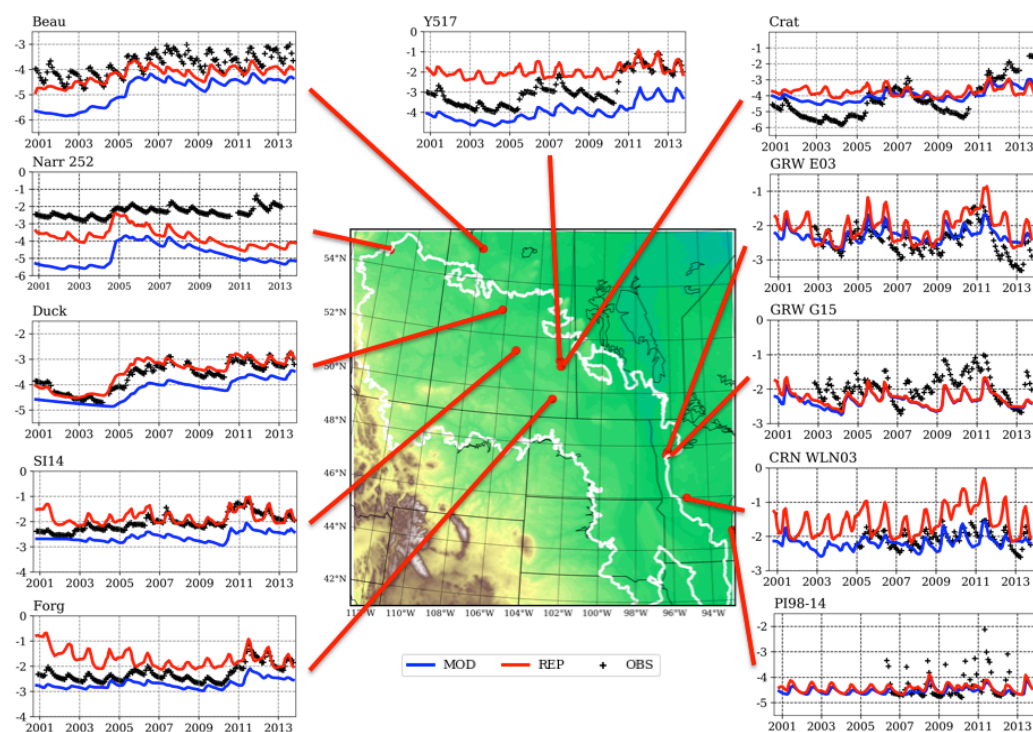
**Fig. 9** Same as Fig. 8, but for the western PPR.

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**Fig. 10** Same as Fig. 6, the timeseries of simulated WTD from both default model (MOD) and replacing soil type simulation (REP). REP is the additional simulation by replacing the default soil type in the model with lithology type taken from observational surveys.