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Modeling groundwater responses to climate change in the Prairie Pothole Region

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Abstract

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

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

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Introduction

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

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

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

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

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

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

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

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

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

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

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

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

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2. Data and Methods

2.1 Observation data

[10] Groundwater observation data were obtained through several agencies: (1) the United States Geological Survey (USGS) National Water Information System in the U.S. (https://waterdata.usgs.gov/nwis/uv?referred_module=gw&search_criteria=search_site_no&search_criteria=site_tp_cd&submitted_form=introduction), (2) the Alberta Environment (http://aep.alberta.ca/water/programs-and-services/groundwater/groundwater-observation-well-network/default.aspx), (3) the Saskatchewan Water Security Agency (https://www.wsask.ca/Water-Info/Ground-Water/Observation-Wells/), and (4) the Manitoba Water Stewardship Division (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, 61 from Saskatchewan, and nine from Manitoba.

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

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

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

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

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

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2.2 Groundwater Scheme in Noah-MP

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

[13] Fig. 2 is a diagram of the structure of Noah-MP soil layers and the MMF aquifer. The active 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. The MMF scheme defines explicitly an unconfined aquifer below the 2-m soil and an auxiliary soil layer stretching to the WTD, which varies in space and time [m]. The thickness of this auxiliary layer (z_{aux} [m]) is also variable, depending on the WTD:

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

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

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

where q is water flux between two adjacent layers [m/s], K_{SMC} is the hydraulic conductivity [m/s] at certain soil moisture content *SMC* [m^3/m^3], Ψ is the soil capillary potential [m] and *b* is soil pore size index. The subscript *SAT* denote saturated state whose values are from the default Noah-MP soil tables. Therefore, the recharge flux from/to the layer above WTD, *R*, can be obtained according to WTD:

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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Fig. 2 Structure of the Noah-MP LSM coupled with MMF groundwater scheme, the top 2-m soil is consist of 4 layers 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 auxiliary layer and the saturated aquifer. Positive flux of *R* denotes downward transport. Two water table are shown, one within the 2-m soil and one below, indicating that the model is capable to deal with shallow as well as deep water table.

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2.3 Forcing Data

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

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

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

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

occurs in summer in the southeast of domain, corresponding to the reduction of future precipitation, as seen in **Fig. 4**.

Fig. 4 Seasonal accumulated precipitation from current climate scenario(CTRL), future climate scenario (PGW) and projected change (PGW-CTRL) in the forcing data.

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

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2.4 Model Setup

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

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

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

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3. Results

3.1 Comparison with groundwater observations

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

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

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

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

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3.2 Climate change signal in Groundwater fluxes

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

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

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

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

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3.3 Water budget analysis

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

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

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

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

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, ΔSMC) and (4) groundwater fluxes and the change of groundwater storage (R, Q_{lat}, Q_r, ΔS_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)-ΔS_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.

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

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

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

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

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4. Discussion

4.1 Simulated WTD sensitivity to soil property parameters

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

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

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

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

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

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

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

4.2 Climate change Impacts on Groundwater Hydrologic Regime

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

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

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

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

Conclusion

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

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

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

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

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

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

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

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

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

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

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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

Table 2. Summary of mean and standard deviation (in the parenthesis) of WTD from 11 groundwater wells, from observation records, default model (MOD) and replacing "true" soil properties simulation (REP). Bold texts show improvement in the mean and seasonal variation of WTD.

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	**-4.20 (0.32)**	**-3.75 (0.51)**	**-3.54** (0.61)	**-1.82** (0.27)	**-1.66 (0.46)**	**1.98** (0.32)	**-3.64** (0.28)	**-1.96 (0.40)**	**-2.26 (0.22)**	**-1.88 (0.43)**	**-4.43 (0.16)**

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(a) (b)

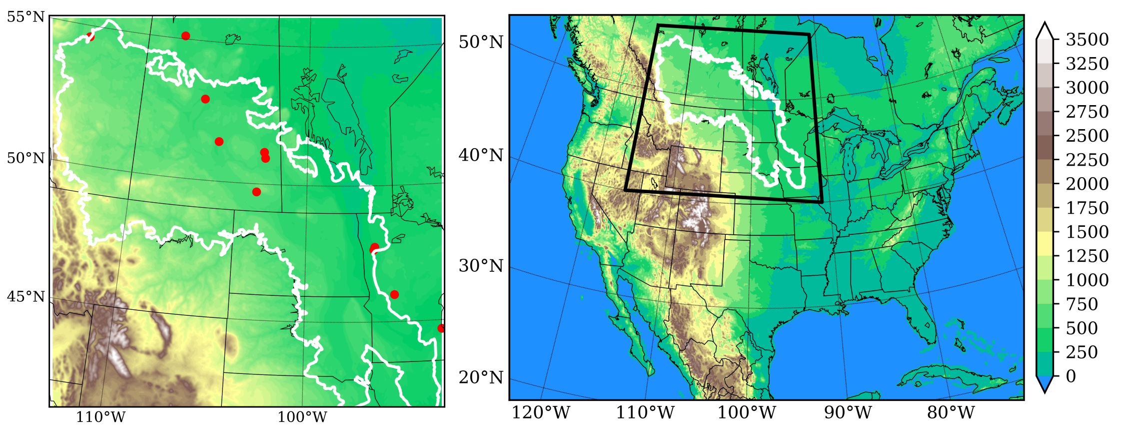

Fig. 1 (a) Topography of the Prairie Pothole Region (PPR; white outline) and station location of rain gauges (gray dots) and groundwater wells (red dots); (b) Topography of the WRF CONUS domain, the black box indicates the PPR domain.

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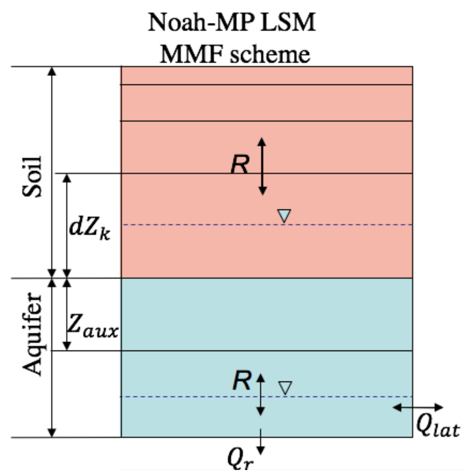

Fig. 2 Structure of the Noah-MP LSM coupled with MMF groundwater scheme, the top 2-m soil is consist of 4 layers whose depth are 0.1, 0.3, 0.6 and 1.0 m. An unconfined aquifer is added below the 2-m boundary, including an auxiliary layer and the saturated aquifer. Positive flux of *R* denotes downward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

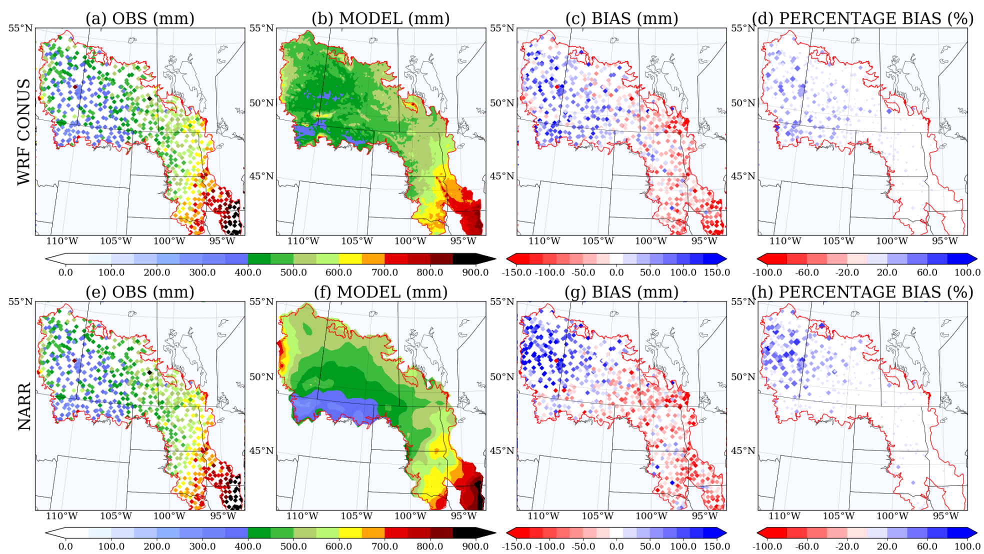

Fig. 3 Evaluation of the annual precipitation from two model products (b, f), WRF CONUS and NARR against rain gauge observation (a, e), their bias (c, g) and percentage bias (d, h).

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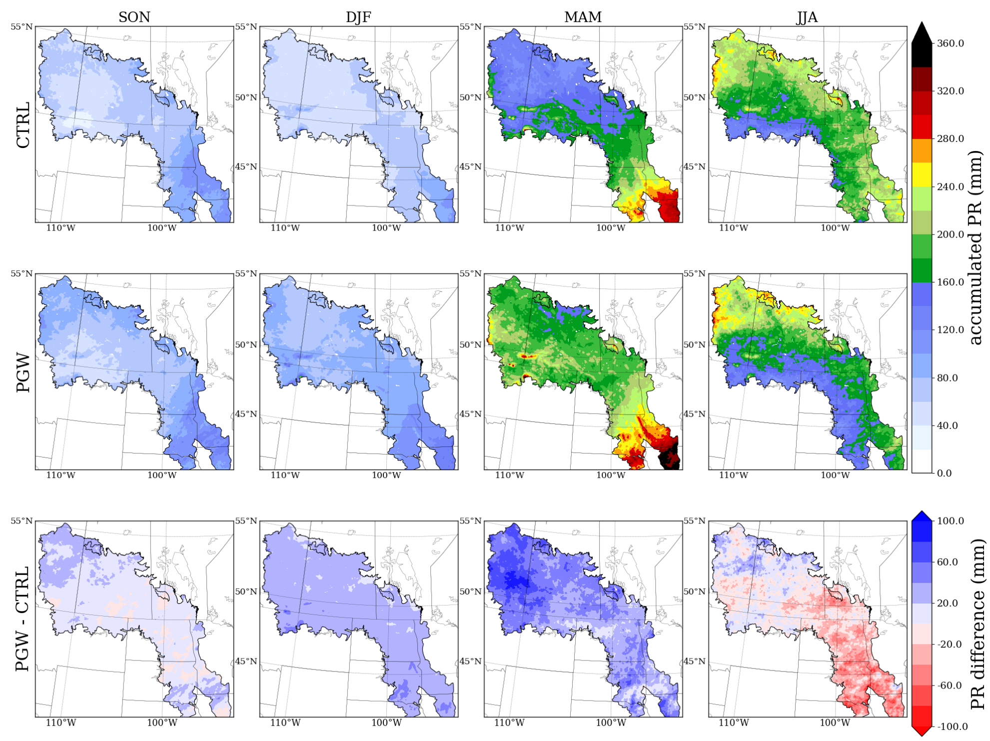

743
744 **Fig. 4** Seasonal Accumulated precipitation from current climate (CTRL, top), future climate (PGW, middle) and
745 projected change (PGW-CTRL, bottom) in forcing data.
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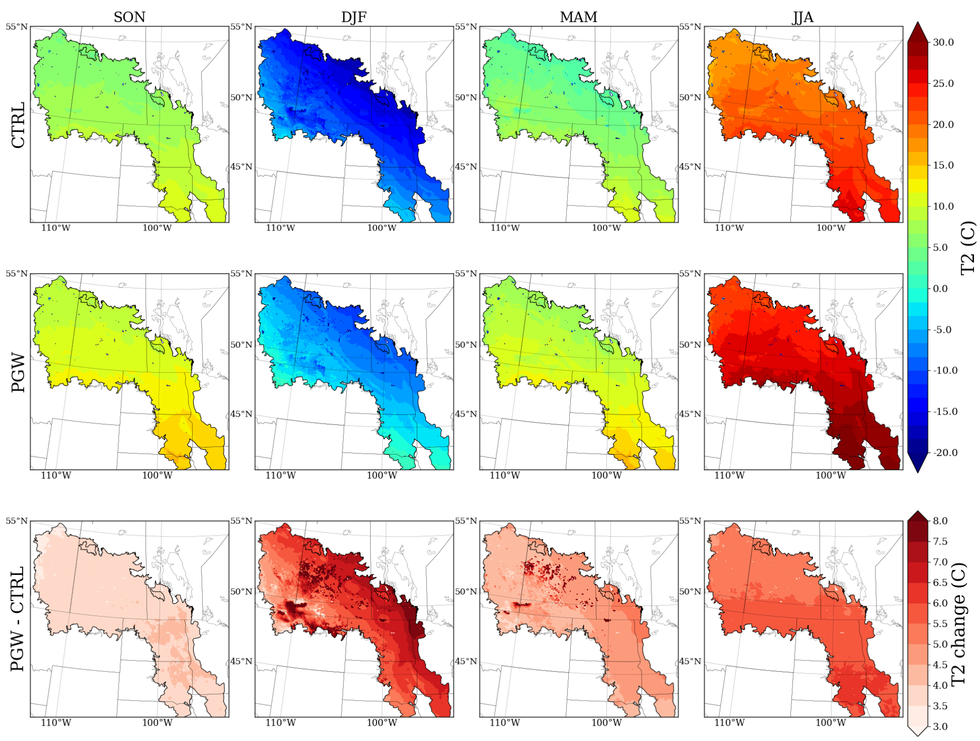

Fig. 5 Seasonal temperatures from current climate (CTRL, top), future climate (PGW, middle) and projected change (PGW-CTRL, bottom) in forcing data.

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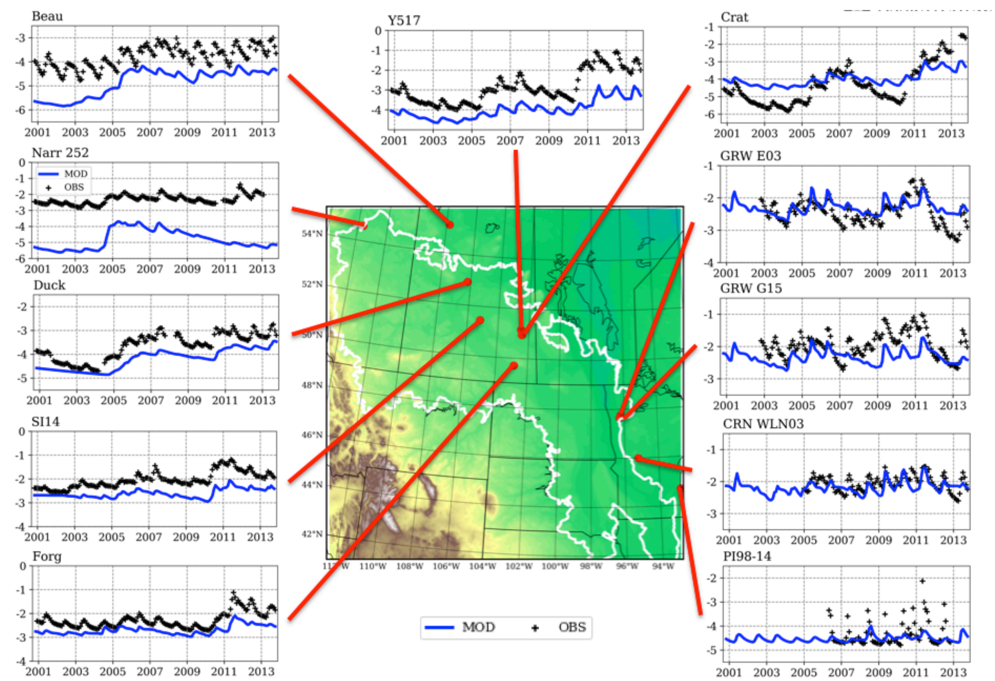

Fig. 6 WTD (m) from 11 groundwater wells and the model simulation results in PPR.

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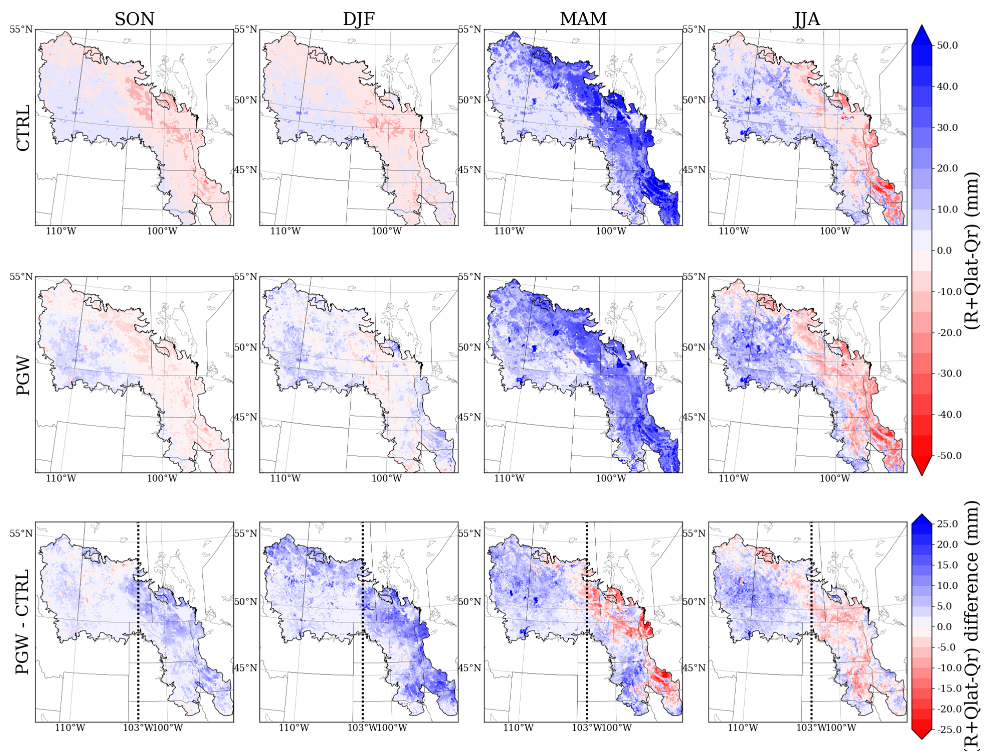

Fig. 7 Seasonal accumulated total groundwater fluxes (R+Q_{lat}+Q_r) for current climate (CTRL, top), future climate (PGW, middle) and projected change (PGW-CTRL, bottom) in forcing data.Black dashed lines in PGW-CTRL separate the PPR into eastern and western halves.

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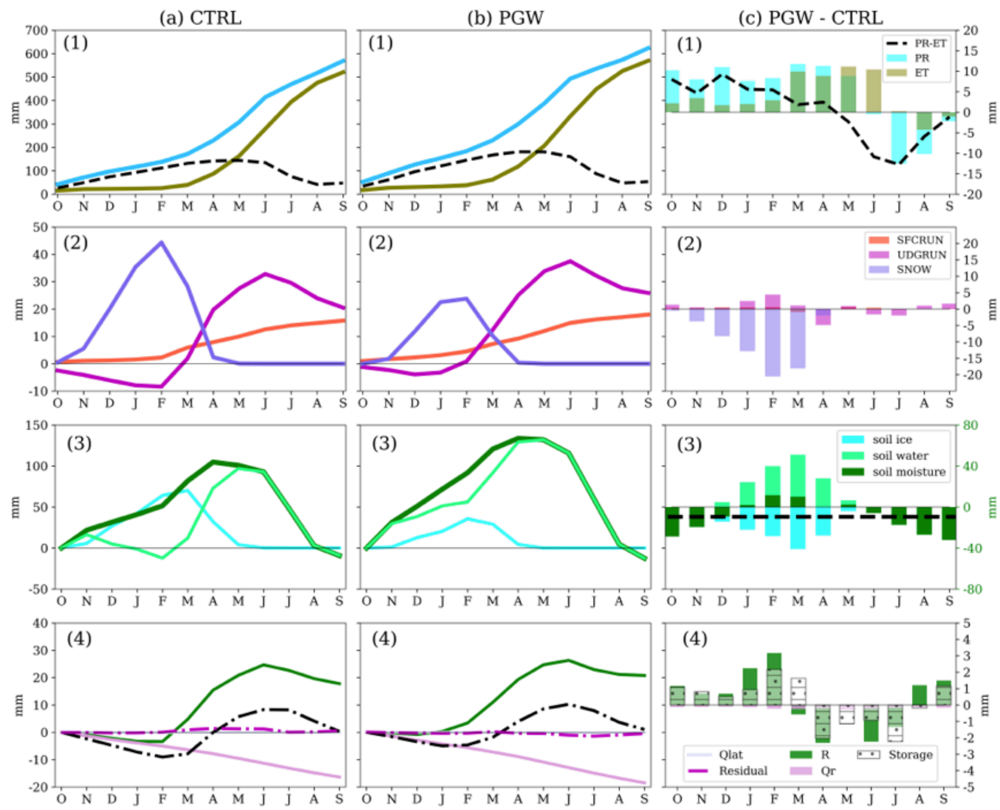


```
storage change:  1.763 mm
recharge change:  4.152 mm
river flux change:  2.260 mm
lateral flux change:  -0.001 mm
```

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, ΔSMC) and (4) groundwater fluxes and the change of groundwater storage (R, Q_{lat}, Q_r, ΔS_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.

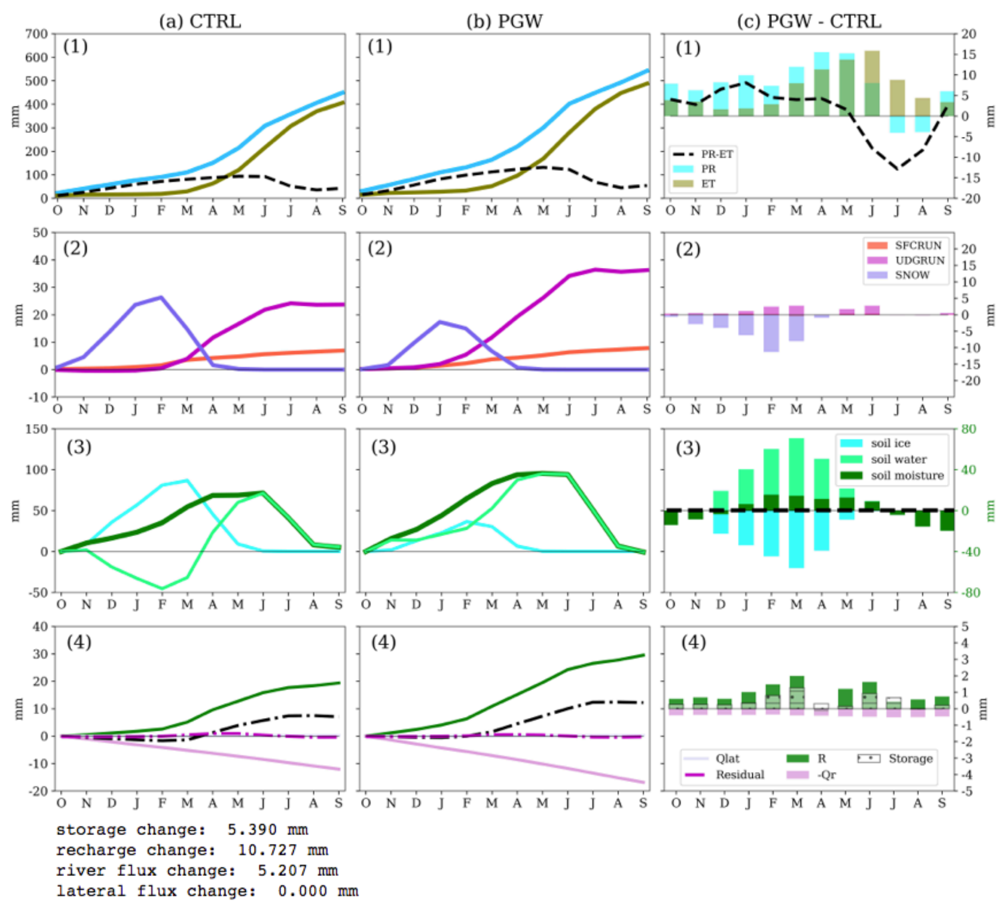

Fig. 9 Same as Fig. 8, but for the **western PPR.**

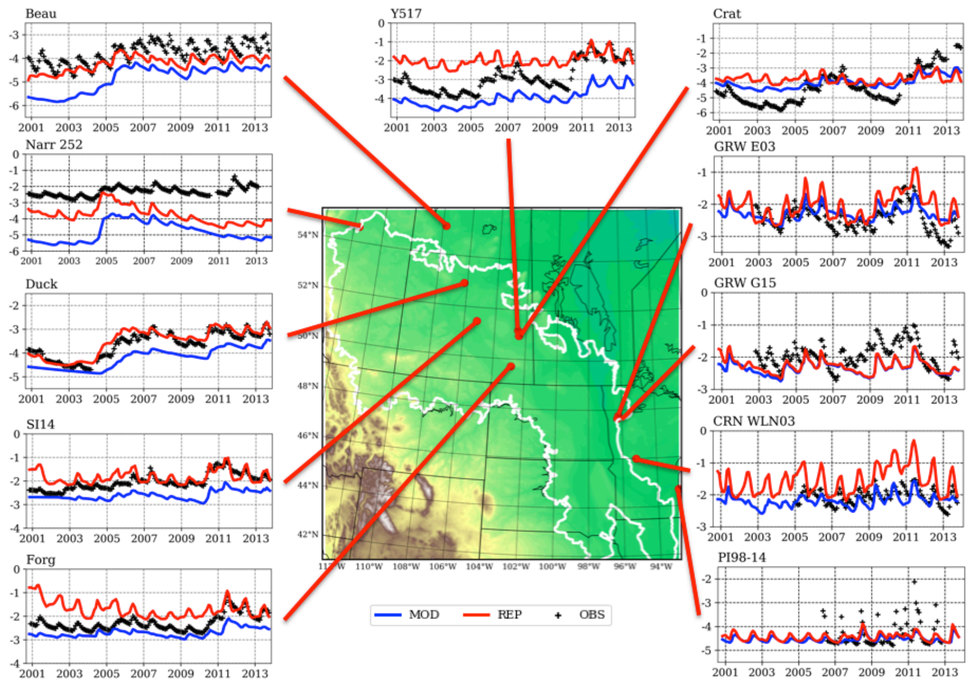

Fig. 10 Same as Fig. 6, the timeseries of simulated WTD from both default model (MOD) and replacing soil type simulation (REP). REP is the additional simulation by replacing the default soil type in the model with lithology type taken from observational surveys.