The Decline In Summer Fallow In The Northern Plains Cooled Near-Surface Climate But Had Minimal Impacts On Precipitation

Gabriel Bromley (bromlgab@gmail.com)
Montana State University Bozeman  https://orcid.org/0000-0002-4497-1058

Andreas F. Prein
National Center for Atmospheric Research

Shannon E. Albeke
University of Wyoming

Paul C. Stoy
University of Wisconsin-Madison

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Abstract

Land management strategies can moderate or intensify the impacts of a warming atmosphere. Since the early 1980s, nearly 116,000 km$^2$ of crop land that was once held in fallow during the summer is now planted in the northern North American Great Plains. To simulate the impacts of this substantial land cover change on regional climate processes, convection-permitting model experiments using the Weather Research and Forecasting (WRF) model were performed to simulate modern and historical amounts of summer fallow, and were extensively validated using multiple observational data products as well as eddy covariance tower observations. Results of these simulations show that the transition from summer fallow to modern land cover lead to ~1.5 °C cooler temperatures and decreased vapor pressure deficit by ~0.15 kPa during the growing season, which is consistent with observed cooling trends. The cooler and wetter land surface with vegetation leads to a shallower planetary boundary layer and lower lifted condensation level, creating conditions more conducive to convective cloud formation and precipitation. Our model simulations however show little widespread evidence of land surface changes effects on precipitation. The observed precipitation increase in this region is more likely related to increased moisture transport by way of the Great Plains Low Level Jet as suggested by the ERA5 reanalysis. Our results demonstrate that land cover change is consistent with observed regional cooling in the northern North American Great Plains but changes in precipitation cannot be explained by land management alone.

Introduction

Global temperatures are rising, primarily due to greenhouse gas emissions from anthropogenic activities (Stocker et al. 2014). Future temperature increases are exceedingly likely (IPCC 2007), as is an increase in precipitation extremes in extratropical zones (O’Gorman and Schneider 2009). Embedded within this global context are changes to regional temperatures and precipitation (Christensen et al. 2007) that often result from the impacts of land management and land cover change on regional energy balances (Mahmood et al. 2014; Luyssaert et al. 2014). Some of these regional climate changes may be beneficial to agricultural and ecosystem management objectives, such as dampening extreme temperatures (e.g. (Juang et al. 2007; Mueller et al. 2015); others are not (e.g. Marshall et al. 2003; Mande et al. 2015). It is important to understand how land management impacts climate processes to develop strategies to minimize the deleterious impacts of climate change and become effective stewards of the earth system.

A unique interaction between land management and climate may have emerged across the northern Great Plains of North America (hereafter the NNAGP). Beginning in the 1960s and 1970s, concerns over soil health and profitability lead to widespread changes in agricultural management away from wheat-fallow rotation agriculture and toward a more diverse agricultural system that avoids bare ground by rotating wheat with pulses, cover crops, and other crops (Miller et al. 2002, 2003; Long et al. 2014). These changes appear to have unintentionally benefitted regional climate (Gameda et al. 2007). Agricultural areas of the Canadian Prairie Provinces have experienced a 6 W m$^{-2}$ cooling during summer across parts of this time period (Betts et al. 2013a). Summer maximum temperatures in the Canadian Prairies have
decreased by nearly 2°C and extreme temperature events have become less frequent (Betts et al. 2013a,b). These regional climate effects have been attributed to the widespread decline of summer fallow from ca. 110,000 km² (25% of Canada’s cultivated lands) to some 35,000 km² (8%) (Gameda et al. 2007; Vick et al. 2016). In the U.S. portion of the NNAGP and across a similar time period, near-surface air temperatures have cooled by nearly 0.2°C decade\(^{-1}\) during late spring and early summer, and near-surface atmospheric vapor pressure deficit (VPD) – which strongly depletes crop yields (López et al. 2021) – has decreased by \(-0.04\) kPa decade\(^{-1}\) on average (Bromley et al. 2020). These changes to the regional climate are concurrent with a period of summer fallow decline on the order of 50,000 km² in the United States from a peak in 1987 until 2012 (Fig. 1).

The observed climate changes are consistent with a transition of large expanses of land away from bare ground and toward crops which actively transport water from soil to atmosphere and increase latent heat flux and evaporative cooling (Vick et al. 2016). Changes to land management have likewise decreased the surface-atmosphere flux of sensible heat to help create a moister, shallower atmospheric boundary layer during summer (Gameda et al., 2007; Vick et al., 2016) through decreases in the Bowen ratio. Combined, these changes in surface fluxes have enhanced cloud formation (Betts et al. 2013b) and increased the probability of convective precipitation (Gerken et al., 2018). Monthly mean precipitation has increased in the Canadian Prairies by 10 mm decade\(^{-1}\) (Betts et al. 2013b; Gameda et al. 2007) and the U.S. northern Great Plains by 8 mm decade\(^{-1}\) (Bromley et al. 2020). Given the multitude of factors that drive precipitation change, the full suite of mechanisms that underlie these observed increases in precipitation and the potential role of land cover change remain uncertain.

Empirical observations and localized modeling studies to date have made critical inroads into our understanding of land-atmosphere-precipitation connections in the NNAGP. Planting crops at the expense of summer fallow decreases planetary boundary layer (PBL) height (Gameda et al. 2007; Vick et al. 2016; Gerken et al. 2018) and, coupled with increases in humidity, lowers the lifted condensation level (LCL) (Betts et al. 2013b; Betts and Desjardins 2018). Shallow cumulus clouds can result when the PBL crosses the LCL, a ‘necessary but not sufficient condition’ for the formation of convective precipitation (Juang et al. 2007). Calculations of PBL and LCL height based on eddy-covariance data and one-dimensional mixed-layer atmospheric models show that the likelihood of PBL-LCL crossings are maximized in May and June when ‘wet coupling’ (Roundy et al. 2013) prevails in the NNAGP such that increased moisture increases the likelihood of convective events (Gerken et al., 2018). This contrasts the prevailing ‘dry coupled’ conditions later in summer when convective precipitation is unlikely (Gerken et al. 2018). Mixed layer models of atmospheric boundary layer development and land-atmosphere coupling have demonstrated an increase in PBL-LCL crossings and an increase in the likelihood of convective rain events across parts of the NNAGP (Gerken et al. 2018), but the mechanisms underlying potential changes in convective precipitation across larger regions have not been explored.

At the same time, changes to surface-atmosphere fluxes due to land cover change may play a minor role in key aspects of the hydroclimate of the NNAGP. Convective precipitation in the NNAGP is dominated by
mesoscale convective systems that are responsible for as much as 60% of the warm season precipitation (Carbone and Tuttle 2008). These systems form in the west and propagate eastward, often overnight, and mixed layer models are generally unable to account for these dynamics (Carbone and Tuttle 2008; Gerken et al. 2018). The buildup of convective available potential energy (CAPE) that supports the development of MCSs comes primarily from the advection of warm, moist air into the region in addition to the diabatic heating of the boundary layer by way of sensible and latent heat fluxes from the surface (Agard and Emanuel 2017). The reduction of summer fallow has the potential to influence the flux of heat and moisture into the boundary layer, leading to the buildup of CAPE. Several studies have posited that increased evapotranspiration from more continuous cropping – and less summer fallow – has led to more growing season convection and potentially stronger storms (Raddatz 1998; Shrestha et al. 2012). However, since the boundary layer has also become cooler and moister, the cooling may act to reduce CAPE and perhaps balance the tendency of added moisture to increase CAPE. Increases in temperature and moisture aloft will also increase convective inhibition (CIN) which may balance or even dominate the effects of any increase in CAPE. The exact response of convective processes to such changes in near-surface conditions is unclear and requires a mechanistic modeling environment that can explicitly account for the dynamics of convective precipitation across regional scales.

How do changes to agricultural management impact regional atmospheric and climate processes? We seek to understand how land management impacts the regional climate and hydrometeorology of the NNAGP, a critical global breadbasket for wheat production. To do so, we model the regional impacts of the reduction of summer fallow across the NNAGP using the Weather Research and Forecasting model (WRF) at a 4 km spatial grid to explicitly model convective precipitation processes across multiple year periods. After validating the model using reanalysis data products and describing the major results of the modeling analysis, we explore reanalysis datasets for a more comprehensive view of the changing hydroclimate of the NNAGP. We then discuss results in the context of the local and large-scale patterns that are consistent with observed climate trends across the region.

**Methods And Experimental Design**

### 2.1 Study Area

We define the semi-arid NNAGP following Bromley et al. (2020) as the combination of the Canadian Prairie Ecozone and the U.S. National Ecological Observation Network (NEON) Domain 9. Briefly, the NNAGP are dominated by grasslands, shrublands, and agriculture with minimal urban development and forests in isolated mountain ranges and river valleys (Stoy et al. 2018). The NNAGP is a critical region for the global production of wheat, pulses, and oilseeds, and corn-soy cropping is becoming increasingly common in its eastern portion (Maaz et al. 2018; Rosenzweig and Schipanski 2019) which, coupled with other land management pressures (e.g. Dolan et al. 2020), create a dynamic system characterized by notable recent changes in land management and widespread increases in vegetation greenness (Brookshire et al. 2020) and decreases in bare ground fraction (Song et al. 2018).
2.1 Model Setup

The Weather Research and Forecasting model (WRF; Skamarock et al. 2008; Powers et al. 2017) is a state of the art weather model that can explicitly simulate convective processes and has been increasingly used in high-resolution regional climate simulations (Powers et al. 2017; Liu et al. 2017; Wang et al. 2018). WRF version 4.1 was run on the Cheyenne computing cluster with a horizontal grid spacing of 4 km, 51 vertical levels up to 50 hPa, and a 20 s time step (Computational And Information Systems Laboratory 2017). The study domain consists of 796 × 496 grid cells and encompasses the NNAGP with at least 40 grid cells between the study area and the edge of the domain. Initial and lateral boundary conditions were provided by the European Center for Medium-range Weather Forecasting’s ERA5 reanalysis at three-hourly intervals (Hersbach et al. 2020). The high resolution of ERA5 (~ 31 km) allows us to directly downscale to 4 km grid spacing without an intermediate nest. The YSU boundary scheme (Hong et al. 2006), Thompson microphysics scheme (Thompson et al. 2008), and the RRTMG radiation scheme (Iacono et al. 2008) were utilized in the simulations. We did not use spectral nudging due to the limited size of the domain.

We used the Noah-MP land surface model with dynamic vegetation options turned off and leaf area index prescribed by table values for each land cover category (dveg = 4) (Niu et al. 2011). Vegetation fraction (fveg) was fixed at the annual maximum to facilitate land cover experiments. Full three-dimensional output was saved every three hours while surface and precipitation data were saved at hourly intervals.

The TOPMODEL groundwater option was turned on, as simulating groundwater and runoff helps reduce the Great Plains warm season bias that is common to many WRF simulations (Barlage et al. 2021). The simulations were run for three years coinciding with the water year beginning in October. The control simulation with lesser summer fallow fraction is simulated for October 2010 – October 2013. It is difficult to avoid the effects of natural climate variability such as those induced by the El Nino Southern Oscillation (ENSO) in short simulations, and results are subject to forcing from ENSO and other climate modes. The simulation period starts during a positive ENSO phase and then shifts to more neutral conditions.

b. Land Cover Experiments

Summer fallow was represented within the model by reducing the fveg parameter by the estimated fallow percentage for each grid cell. Noah-MP uses a split-cell calculation for land-atmosphere interactions, meaning fluxes are calculated for the bare ground fraction and the vegetated fraction separately and then combined to give the surface fluxes for the entire grid cell. The control simulation uses a fveg that matches the estimated summer fallow extent in 2011 (hereafter ‘C11’), while the simulation with modern climate and 1980s summer fallow extent is called F11. We chose 2011 to match data availability from the U.S. National Land Cover Dataset (NLCD) (Homer et al. 2015), noting that this year was subject to relatively large fallow areas in parts of Saskatchewan, Manitoba, North Dakota and Minnesota due in part to widespread spring flooding that limited planting (Fig. 1) (Stadnyk et al. 2016). From this perspective,
our analysis represents a conservative interpretation of fallow change from the 1980s until the 2010s. We focus our analyses on the C11 and F11 simulations to isolate the role of land cover change apart from decadal global climate change on determining regional climate changes in the NNAGP. Statistical differences between simulations were assessed using the Mann-Whitney U test on daily data. The data processing workflow relied on the Climate Data Operators package (Schulzweida 2019) as well as several analysis packages in Python, such as Numpy, Xarray, Matplotlib, and Scipy Stats (Harris et al. 2020a; Hoyer and Hamman 2017; Hunter 2007; Virtanen et al. 2020). Convective parameters, such as CAPE and convective inhibition (CIN) were calculated using the cape_2d function from the wrf-python package (Ladwig 2017).

Results

3.1 Model Validation

The control simulations are extensively validated against observational datasets in the Appendix and validation will only briefly be discussed for completeness. Control simulations were compared to the Daymet dataset (Thornton et al. 2016), the CRU dataset (Harris et al. 2020), the Grided Meteorological Ensemble Tool (Newman et al. 2015), and the Global Precipitation Climatology Centre precipitation dataset (Rustemeier et al. 2020). Surface fluxes were compared against eddy covariance observations from Lethbridge, Alberta (Flanagan et al. 2002). The near-surface (2 m) temperature in the C11 simulation was well simulated during spring and fall with cold (warm) biases in Winter (Summer) that were similar in magnitude to other WRF simulations (Liu et al. 2017) in the NNAGP (Figure A2). Precipitation was well simulated in C11 for all seasons (Figure A3). The control simulation was compared to observational atmospheric soundings acquired from the Integrated Global Radiosonde Archive (Durre et al. 2006) for the locations of Edmonton, AB, Glasgow, MT, and Bismark, ND. C11 has a 2.5°C near surface warm bias at all locations for the late warm season, while the early warm season was well captured and was within 1°C at all levels (Figure A7). Surface energy fluxes largely matched eddy covariance measurements from a grassland site in Lethbridge, AB (CA-Let, (Flanagan et al. 2002) (Figure A8).

Investigating warm season climate trends is the focus of this work, and changes to other seasons are intermittently discussed for completeness. Seasonal changes are taken to be the average change across the three-year simulation.

3.2 Changes to summer fallow extent and representation in WRF

The extent of summer fallow in Provinces and States that intersect the study area decreased from 151,900 km² in 1982 to 35,100 km² in 2012, the years closest to the study periods when data were available from the United States (data from Canada were available every year) (Fig. 1). Some 45% of the total decline in fallow of 116,800 km² was attributable to Saskatchewan alone (53,000 km²). These published agricultural statistics were used to nudge the Landsat fallow attribution analysis on a per-
Province and per-State basis (Fig. 2) to ensure that it simulated total fallow area, which was then used to adjust the bare ground fraction in Noah-MP as noted.

### 3.3 Changes to near-surface temperature, energy fluxes, and humidity

Two-meter air temperature (T2) shows a domain-averaged increase of about 0.18°C during the growing season in F11 relative to C11 noting that the study period includes areas that experienced both increases and decreases in summer fallow (Figs. 1–3). The strongest simulated warming is limited to June, July, August (JJA) (Fig. 4). T2 cooled on average by −0.5°C in areas where fallow decreased in C11 compared to F11. The northern and western part of the NNAGP in Alberta experienced a T2 increase on the order of +1.5°C. T2 warmed on average by about 1.0°C in areas where fallow increased in F11 compared with C11 across the entire study domain. There is a linear relationship between fveg and T2 (Fig. 3b); T2 increases by 0.69°C for every 10% increase in summer fallow. There is a modest cooling signal during winter indicating the more vegetated C11 simulation is warmer by 0.25°C (Fig. 4) that follows the same spatial pattern as growing season temperature changes, but opposite in sign and of a lesser magnitude.

| 1980s meteorology ('84') | 2010s meteorology ('11') |
|--------------------------|--------------------------|
| 1980s fallow             |                          |
| C84:                     | F11:                     |
| 1980s meteorology + 1980s fallow estimate | 2010s meteorology + 1980s fallow estimate |
| 2010s fallow             |                          |
| F84:                     | C11:                     |
| 1980s meteorology + 2010s fallow estimate | 2010s meteorology + 2010s fallow estimate |

Changes in VPD between F11 and C11 follow the same spatial pattern as T2; the near-surface atmosphere in areas that increased in fveg became moister and the areas that decreased in fveg became drier (Fig. 5b). The domain median change is −0.045 kPa and with a wide distribution that encompasses both increases and decreases in VPD (Fig. 5a). The most widespread increase in VPD is in Alberta with a 0.15 kPa increase from F11 to C11 over most of the Province within the study area. The strongest increase in VPD is nearly 0.3 kPa and occurs in central and eastern South Dakota. The largest decrease in fveg from F11 to C11 occurs in Manitoba where there was an observed increase in summer fallow (Fig. 1); VPD is lower there by −0.15 kPa, with areas that decrease nearly −0.2 kPa.

Study area-averaged sensible heat flux (H) is higher by 10 W m\(^{-2}\) in F11 compared to C11 in areas where fveg is lower compared to C11 (Fig. 6a). The areas of largest change have magnitudes on the order of 20 W m\(^{-2}\). Changes to latent heat flux (LE) follow the same pattern as H, but the magnitudes are larger (Fig. 6b). Study area-averaged LE is −16 W m\(^{-2}\) lower in the F11 simulation compared to the C11 simulation.
The differences in LE are larger than H, with some areas in excess of 30 W m$^{-2}$ in C11. Eastern Montana, western South Dakota and southern Saskatchewan show smaller magnitudes of change in both H and LE compared to the areas near the border of the study area that are considered ‘crop’ land cover types in the land surface model.

### 3.4 Changes to convective environments

We separate our analyses of convective environments between the early warm season (May and June) and late warm season (July and August) given differences in surface-atmosphere coupling in the NNAGP during these periods (Gerken et al., 2018).

During May and June, changes to CAPE and CIN are not significantly different from background noise (Fig. 7a,b). The height of the LCL is higher in F11 compared to C11 for most of the study area. The mean change to LCL height is 13 m while some areas are more than 30 m (Fig. 7c). Manitoba is the only area where the LCL heights have decreased. Mean LCL heights in Manitoba decreased by 20–30 m. Differences in planetary boundary layer (PBL) heights closely follow the spatial pattern of differences in LCL heights (Fig. 7d) and, as a consequence, the areas that have a positive change in fveg from F11 to C11 have higher PBL heights, while the opposite holds for areas that have a negative change in fveg such as Manitoba. The mean change in PBL heights within the study area is 8 m but some areas change up to 30 m.

During July and August, CAPE is lower across most of the study area in the F11 simulation by $-10$ J kg$^{-1}$ with minima in Alberta and the Dakotas (Fig. 7a) where CAPE decreases by 20–40 J kg$^{-1}$. Changes to CIN in July and August weaker than the changes to CAPE (Fig. 8b). CIN is lower in the F11 simulation than the C11 simulation in Alberta and parts of Saskatchewan by over $-5$ J kg$^{-1}$ on average. Manitoba and north-eastern North Dakota exhibit higher CIN in F11 than C11 by 20–30 J kg$^{-1}$. Differences in LCL heights show a similar pattern to May and June but much stronger with F11 simulating LCL heights that are 50 m higher than C11 across most of the study area (Fig. 8c). The largest differences are located along the North Dakota-South Dakota border, northern Saskatchewan, and Alberta. LCL heights in these areas are over 100 m higher. PBL heights in July and August follow a similar spatial pattern as in May and June but with greater magnitude (Fig. 8d). PBL heights in Alberta and northern Saskatchewan are higher in the F11 simulation than C11 by over 100 m on average but only 10-20m higher in central parts of the study area are. The only area where PBL heights are lower in F11 is in Manitoba with an average change of about 50 m, noting again that there was an increase in simulated fallow extent in Manitoba between 1984 and 2011 (Figs. 1 & 2).

### 3.5 Changes to precipitation

Changes to precipitation were not appreciably different from background noise from May through August (Fig. 9). When aggregated to 100 km a slight drying becomes apparent during July and August, but any effects are not statistically significant.
Discussion

We demonstrate using convection-permitting WRF model simulations that land management change toward continuous cropping and away from summer fallow decreased near-surface air temperature and VPD but had muted impacts on precipitation in part because increases in instability through increased boundary layer moisture have been balanced by an increase in stability through a cooler boundary layer. Precipitation did not appreciably change between the WRF simulations, indicating a possibility that precipitation processes in the NNAGP are not very sensitive to the land surface changes of the magnitude experienced in recent decades. Below, we elaborate on each of these findings to describe how land cover change has modified important aspects of the regional climate of the NNAGP, especially near the land surface. We then add to emerging evidence that observed changes in precipitation are likely due to moisture advection into our study domain rather than regional surface-atmosphere interactions.

Temperature

To summarize findings on the impact of summer fallow on near-surface climate: areas that underwent a fveg increase from F11 to C11 were cooler with lower VPD (Figs. 3 & 6). Near-surface warming and drying occurred in areas where fveg decreased. These results lend evidence to the notion that a reduction in summer fallow is largely responsible for the cooling and moistening trend that is observed across the NNAGP. The changes to temperature in the simulations are stronger than the trends calculated by Bromley et al., (2020), noting that the trends in the latter are calculated from a 1970 starting point, whereas these experiments simulate fallow reduction from the 1980s to 2010s. Temperature trends are stronger at nearly $-0.5^\circ$C decade$^{-1}$ (Bromley et al. 2020) when calculated using 1980 as a starting point, on the order of 1–1.5$^\circ$C, similar to modeled changes in T2 associated with an fveg increase from F11 to C11. The temperature difference simulated here spans from May until September, but given wheat is often harvested in August (if not sooner for the case of winter wheat), the September T2 difference is likely due to the prescribed seasonal cycle for each land use category.

The winter warming in the C11 simulation relative to the F11 simulation is likely due to the decrease in albedo from increased fveg in the model. Since the fveg does not change based on a seasonal cycle, areas with greater fveg are assumed to be lower in albedo since the vegetation is not covered in snow. The bare ground areas are covered in snow and thus are higher in albedo. This is similar to year-round cover cropping and the winter warming effect has been noted in global climate models (Lombardozzi et al. 2018). Simulating snow advection, especially the tendency of vegetation to trap blowing snow (Pomeroy and Li 2000; Pomeroy et al. 1998), is a challenge to earth system models, which may overestimate the albedo effect of wintertime vegetation as a consequence which (Hunter et al. 2019) Multiple modifications to the Noah-MP snow physics calculations were made by Liu et al., (2017), to create more realistic cold season surface-atmosphere interactions and spring melt profiles but wintertime processes are still an active area of land surface model research.

Vapor Pressure Deficit
Plant stomata respond strongly to VPD. If VPD is too high, stomata will close to avoid evaporative water losses, effectively shutting off carbon uptake by plants (Eamus et al. 2013; Grossiord et al. 2020; Novick et al. 2016) with important implications for surface-atmosphere fluxes (Yuan et al. 2019; Rigden and Salvucci 2017). VPD is increasing on average across the United States, except for the U.S. portions of the NNAGP which are decreasing in VPD by an average of 0.5 kPa decade$^{-1}$ (Bromley et al. 2020; Ficklin and Novick 2017). The VPD change for an increase in fveg from F11 to C11 is on the order of $-0.45$ kPa which corresponds to a first order with the observed changes to VPD in the NNAGP.

Our modeling analysis suggests that the impacts of simulated fallow reduction on near surface climate has acted to create more favorable conditions for crop growth by reducing growing season temperatures and VPD (Hsiao et al. 2019). Wheat yields differ in their sensitivity at different crop growth stages, and early-season days with mean temperature $>28^\circ$C are especially detrimental (Asseng et al. 2015). It is interesting to note that the midwestern United States has experienced largely beneficial changes in near surface growing season climate as a result of agricultural intensification (Mueller et al. 2015), leading one to question if they can be sustained in the future as global climate change continues to stress water resources and production systems.

**Boundary layer changes**

The PBL by definition is the near surface layer of the atmosphere that is strongly influenced by surface fluxes of water and energy, so it is not surprising that the systematic shift away from summer fallow affects PBL processes. The monthly mean boundary layer heights were 100 m higher in late summer in the F11 simulation where the fallow amounts were larger; the lowering of the PBL as fallow declines was proposed to be a consequence of the changes in energy partitioning from a fallow (bare) surface and a vegetated surface (Gameda et al. 2007) which was the case in these simulations (Fig. 5). The change in PBL height was previously assessed using a simple slab model with inputs from eddy covariance observations of turbulent fluxes from wheat and fallow fields (Vick et al., 2016). These simulations suggested an increase in PBL height of about 200 m during the growing season over a fallow field versus a spring wheat field. This 200 m difference is larger than the 60 m difference in mean monthly PBL heights simulated here, which is perhaps not surprising given that WRF simulates a spatial mix of fallow and vegetated areas whereas Vick et al. (2016) modeled PBL impacts of fallow and vegetation separately. PBL growth is sensitive to heterogeneous landscapes and the model representation of seasonal and diurnal variations is improved if the heterogeneity of surface fluxes are captured (Rey-Sanchez et al. 2021). Using a model similar to WRF, MM5, (Mahmood et al. 2011) found that simulations of bare soil were 1.4°C warmer than the control simulations (present day vegetation) and the seven-day average PBL heights were $\sim 550$ m higher with a lower LCL under higher fveg fractions, which increased the probability of cloud development and convection (Mahmood et al. 2011). Our results are consistent with the notion that summer fallow changes PBL and LCL heights but its realized impact on precipitation was relatively small and spatially variable (Fig. 9).

**Precipitation**
There is little evidence that precipitation changed appreciably between F11 and C11 (Fig. 9), suggesting that a reduction in summer fallow might not have had much of an impact on observed precipitation trends. July and August are 10–15 mm drier in the F11 simulation when the precipitation change is aggregated to 100 km × 100 km boxes but these changes are not significant at the 95% level. Precipitation in the NNAGP increased by 8 mm decade$^{-1}$ in May and June, but July and August precipitation also increased, primarily on the eastern side of the NNAGP (Bromley et al. 2020). If the land surface is not appreciably changing mean precipitation, what is the source for the observed warm season precipitation increase (Bromley et al. 2020)?

Global mean precipitation has been increasing due to anthropogenic warming of the atmosphere at about a rate of 2% K$^{-1}$ (Held and Soden 2006; Pendergrass and Hartmann 2014). This rate comes from the thermodynamic change to precipitation, but does not account for changes to the dynamic components such as changes to circulation. Precipitation in the NNAGP is largest in the early warm season and May through September is a convectively-active period (Gerken et al. 2018). Precipitation during this time period can take the form of stratiform rain, MCSs, and organized pre-frontal convection; July and August are quite dry compared to May and June and precipitation is primarily from MCSs. The Great Plains Low Level Jet (GPLLJ) is a nocturnal wind speed maxima, positioned at about 850 hPa, that transports moisture from the Gulf of Mexico into the Great Plains, and July and August MCS development in the NNAGP is usually accompanied by a strong northward-penetrating GPLLJ (Feng et al. 2019; Song et al. 2019; Feng et al. 2016). To investigate the possibility that the observed increase in precipitation in the NNAGP is consistent with additional moisture sources from the south, we investigated meridional wind and specific humidity trends in ERA5. Figure 10a shows a vertical cross-section along the 42º latitude line of 1979–2020 trends in monthly mean meridional wind and specific humidity. Due to the lack of strong trends in meridional wind, meridional moisture transport trends are only slightly positive (Fig. 10b). There is not a clear signature of a strengthening GPLLJ, but the increase in surface specific humidity corroborates Bromley et al., (2020); near-surface conditions are moistening during May and June. Trends in the North American Regional Reanalysis (NARR) dataset shows that moisture transport northward has increased during AMJ, particularly during days with MCS initiation (Feng et al. 2019; Barandiaran et al. 2013).

Most of the MCSs in the NNAGP occur during July and August, and the northward extension of the GPLLJ over the past four decades is clear in monthly mean trends (Fig. 11). Specific humidity has increased at 0.3 g kg$^{-1}$ decade$^{-1}$ while meridional wind has increased at 0.35 m s$^{-1}$ decade$^{-1}$. These trends add moisture to the NNAGP and likely contribute to the observed increase in precipitation on the eastern and southern boundaries of the NNAGP during summer as well as the lower VPD during JJA (Bromley et al., 2020). The magnitude of CAPE change was on average larger in July and August, which could mean that the increase in convective environments conducive to strong storms has been aided by the reduction of fallow (Brimelow et al. 2011). An analysis that tracks MCSs and looks at changes to convective environments, e.g. Feng et al, (2016), could perhaps show how much the land surface impacts
these processes and resolve how changes in land cover and regional circulation processes have impacted the unique climate trends of the NNAGP.

**Summary And Conclusions**

Summer fallow in the NNAGP has declined from an estimated 151,900 km\(^2\) in the 1980s to 35,100 km\(^2\) in the 2010s, a decline of 116,800 km\(^2\) which is approximately the land area of Pennsylvania. To investigate the climate impacts of this reduction in summer fallow, two three-year convection-permitting WRF simulations were performed, using ERA5 as the initial and lateral boundary conditions. The vegetation fraction of each simulation was adjusted using Landsat-estimated summer fallow and nudged to match published agricultural statistics for 2011 and 1984. The intention of these simulations is to understand how the near surface climate and precipitation processes have been impacted by these substantial changes in land cover. The summary of the results are:

- Two-meter air temperatures were 1-1.5ºC cooler and VPD was 0.15 kPa lower in areas where \(f_{\text{veg}}\) increased between the fallow simulation and the control simulation.
- The PBL and LCL were lower by 60 m, due to the cooler and more humid land surface.
- CAPE increased by 20–30 J kg\(^{-1}\) but there were minimal changes to CIN.
- Precipitation did not change appreciably between the simulations, but the fallow simulation was 10–15 mm drier during July and August.

The results of these simulations suggest that observed near-surface cooling and moistening trends in the NNAGP are largely a result of the reduction in summer fallow. The lack of evidence for a land-surface induced change to precipitation stands in contrast to other observational studies focused on the same region; however, this is the first modeling study looking at summer fallow reduction. Further work is needed to better understand the precipitation processes, perhaps tracking the evolution of precipitating storm systems as they move over the heterogeneous and changing landscape of the NNAGP.

**Declarations**

**Funding**

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Conflicts of interest
The authors report no conflicts of interest.

Availability of data and material
Monthly mean datasets will be stored in a public repository and available for public use with the author's consent.

Code availability
Code used in the analysis will be available from https://github.com/gbromley.

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Figures

Figure 1

The area of land held in summer fallow in the (A) Canadian Prairie Provinces and (B) U.S. States of the northern North American Great Plains for the 1982 - 2012 period using data from Statistics Canada and the United States Department of Agriculture Economic Research Service following Vick et al. (2016). The primary study years, 1984 and 2011, are indicated with vertical dotted lines.
Figure 2

Differences in vegetation fraction ($f_{veg}$) between the 2010 fallow and 1984 for the Noah-MP land surface model in the Weather Research and Forecasting (WRF) model estimated using Landsat and adjusted to match published agricultural statistics (Fig. 1).
Figure 3

Two-meter temperature (T2) differences between modern (2011) fallow (F11) and control (C11) simulations averaged across the three-year simulations for MJJA. Stippling indicates significant differences at the 95% level.
Figure 4

Monthly differences in T2 between the F11 and C11 WRF simulations (Table 1). Positive values indicate the F11 simulation was warmer.
Figure 5

Two-meter vapor pressure deficit difference between the modern fallow (F11) and control (C11) simulations during the three year simulation period for MJJA. Positive VPD values indicate that the F11 simulation is drier, while negative values indicate that the F11 simulation is moister. Stippling indicates significant differences at the 95% level.
Figure 6

The difference in sensible (a) and latent (b) heat flux for MJJA between the F11 simulation and the C11 simulation. Stippling indicates significant differences at the 95% level.
Figure 7

Changes to (a) convective available potential energy (CAPE), (b) convective inhibition (CIN), (c) lifted condensation level (LCL), and (d) planetary boundary layer (PBL) height between 2011 control (C11) and fallow (F11) WRF simulations during May and June for the three-year simulation period. Stippling indicates significant differences at the 95% level.
Figure 8

Same as Figure 7 but for July and August for the three-year simulation period.
Figure 9

Changes to precipitation for May and June ("Early Warm", Row 1) and July and August ("Late Warm", Row 2) for the F11 and C11 WRF simulations (Table 1), their absolute difference (Abs Diff), and percent difference. Precipitation was aggregated to 100 km × 100 km boxes to display regional trends.
Figure 10

Vertical cross-section of 1979-2020 May and June meridional wind trends (black contours) and specific humidity trends (filled contours) for the levels between 925 hPa to 800 hPa from the ERA5 reanalysis. Inset axes show trends in meridional moisture transport (qv) for 1979-2020 and the location of the cross-section. Brown contour shows the pressure where topography is located along the cross section.
Figure 11

Same as Figure 10 but for July and August.

Supplementary Files

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- Appendix.docx