Warming/cooling effects of cropland greenness changes during 1982–2006 in the North China Plain

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Abstract

This study analysed the changes in cropland greenness during 1982–2006 in the North China Plain (NCP) and investigated the warming/cooling effects of the greenness changes. The results show that while spring cropland greenness increased, early summer cropland greenness substantially decreased from 1982 to 2006. In contrast to the cooling and wetting effects of the greenness increase in spring, the greenness reduction in early summer had warming and drying effects. The cooling/warming effects of cropland greenness changes accounted for ∼47% of the spatial variance of daily maximum temperature ($T_{\text{max}}$) change in spring and ∼44% in early summer. The wetting/drying effects of cropland greenness changes accounted for ∼48% of the spatial variance of daily minimum specific humidity ($\text{SPH}_{\min}$) change in spring and ∼19% in early summer. The cooling–wetting/warming–drying effects mainly resulted from the distinct partitioning of surface net radiation between surface latent heat flux and sensible heat flux over cropland with different greenness. Canopy transpiration plays a dominant role. The increased (decreased) cropland greenness corresponds to high (low) transpiration rate, less (more) sensible heat flux and high (low) humidity, and consequently cooling–wetting (warming–drying) effects. In comparison, there was little change in surface net radiation, although surface albedo and emissivity had changed with greenness change.

Keywords: NDVI changes, warming/cooling effects, North China Plain, canopy transpiration

Online supplementary data available from stacks.iop.org/ERL/8/024038/mmedia

1. Introduction

Terrestrial vegetation is an important, dynamic component of the climate system (e.g., Foley et al. 2005, Bonan 2008, Pielke and Niyogi 2009, Yi et al. 2010, Arneth et al. 2010). There is a growing body of scientific evidence showing that it influences local and regional climate change (e.g., Mahmood et al. 2010, Dirmeyer et al. 2010). Terrestrial vegetation can affect the surface energy balance by changing surface albedo and the partitioning of land surface net radiation between the sensible, latent and ground heat fluxes (e.g., Notaro et al. 2006, Pielke et al. 2007, Anderson-Teixeira et al. 2012), or by changing surface roughness to regulate the atmospheric dynamics and thus the transportation of moisture and heat (e.g., Takata et al. 2009). Therefore, knowledge of the effects of terrestrial vegetation change on climate is crucial for understanding regional climate changes.

The last two decades of the 20th century were characterized by a widespread increase in surface air temperature, an advance of spring vegetation phenology and an increase in surface greenness, as indicated by the
normalized difference vegetation index (NDVI), across the mid-high latitudes of the northern hemisphere (e.g., Zhou et al 2001, Zhou et al 2003, Piao et al 2011, Wang et al 2011b). Human agricultural practices such as irrigation and fertilization may further increase surface greenness in agricultural areas (e.g., Piao et al 2003, Fang et al 2004). Higher surface greenness generally implies higher absorption of photosynthetically active radiation (PAR) (e.g., Myneni et al 1995), more leaves (Baret and Guyot 1991), a higher fraction of green vegetation (e.g., Gutman and Ignatov 1998), and more biomass (e.g., Muñoz et al 2010). These changes in land surface properties may alter the land surface radiation budget and the energy partitioning between sensible heat and latent heat (e.g., Tang et al 2012) and have consequent climatic implications (e.g., Liu et al 2006, Jeong et al 2009a, Wang et al 2011a).

In China, the North China Plain (NCP) is an important agricultural area. The dominant land use type in the NCP is cropland. The prevailing crop system is double harvests per year consisting of winter wheat and summer maize. The seasonal cycle of surface greenness is characterized by two peaks in early May and late August corresponding to maximum growth of winter wheat and summer maize, respectively, and one valley (late May to early June) corresponding to mow-killing of winter wheat (e.g., Xin et al 2002, Lei and Yang 2010, Pan et al 2012). Despite global climate warming the surface greenness during the growing season (from April to October) had no detectable increases in North China, likely because of precipitation changes in the last decades of the 20th century (Xu et al 2012). However, spring surface greenness increased extensively across the NCP, likely because of an increase in spring temperature and improved agricultural practices, including irrigation and fertilization (e.g., Piao et al 2003, Peng et al 2011). Meanwhile, the cropping system also shifted (Zhang and Xu 2010). In the early 1980s, an interplanting technique was practised in which maize was planted before the mow-killing of winter wheat (Tong 1979). When the winter wheat was moved in late May, the maize had already sprouted. Since the early 1990s, the interplanting has been gradually replaced by successive planting, i.e., planting maize after mow-killing the winter wheat (Zhang and Xu 2010). Under this system, when winter wheat was mowed, maize had not yet sprouted and the bare ground was directly exposed to the atmosphere. With the practice of successive planting, early summer cropland surface greenness would be lower than that with the interplanting technique.

Ground-level measurements in the NCP illustrate that other surface parameters, such as surface albedo, leaf area index (LAI) and Bowen ratio, also have intra-annual variations similar to the seasonal variations in surface greenness. Corresponding to a high phase of NDVI in May and August, surface albedo reaches a low phase; thus, surface net radiation reaches a relative high phase (e.g., Xu and Levy 2011), while the LAI reaches high phase, evapotranspiration has a high value, and surface sensible heat flux has a low value (e.g., Lei and Yang 2010). Corresponding to a low value phase of NDVI in June, the conditions are reversed. Low albedo results in high surface net radiation, and therefore has a potential warming effect, whereas high evapotranspiration and low sensible heat flux have a potential cooling effect. Jeong et al (2009a) reported a negative spatial correlation between climate warming and spring NDVI increase from 1982 to 2000 across East Asia, including the NCP, and noted a cooling effect from increased NDVI. Simulations with the coupled CAM3–DGVM model indicated that the cooling effect dominatedly resulted from evapotranspiration, whereas the warming effect of albedo change was minor (Jeong et al 2009b). However, using a temporal variation method to analyse observed temperature and NDVI, Wu et al (2011) reported that changes in NDVI had no significant effects on NCP spring temperatures from 1982 to 2006. Moreover, we know little about the climatic effects of early summer surface greenness changes.

This study aimed to investigate the effects of changes in the spring and early summer surface greenness on surface air temperature across the cropland area in the NCP from 1982 to 2006. The effects of surface greenness change on temperature were analysed using NDVI data and ground meteorological measurements. The possible mechanisms by which surface greenness changes affect the temperature were investigated using the weather research and forecast (WRF) model. The surface greenness increased in the spring, along with climate warming, but decreased in early summer because of the change in cropping system in the NCP. Thus, we focused on the climatic effects of the spring increase and early summer decrease in surface greenness.

2. Data and methods

2.1. Data

The NDVI data was obtained from the Global Inventory Monitoring and Modeling Studies (GIMMS) group (www.landcover.org/data/gimms/). The GIMMS NDVI was derived from the NOAA/AVHRR Land dataset. The spatial and temporal resolution of these data are 8 km × 8 km at 15 day intervals for the period from January 1982 to December 2006 (Tucker et al 2001, Zhou et al 2001). The NDVI data have been verified and widely used to investigate the vegetation dynamics in a large number of studies in China (e.g., Fang et al 2004, Piao et al 2011, Xu et al 2012).

The daily meteorological observations, including surface air daily mean \( T_{\text{mean}} \), maximum \( T_{\text{max}} \), and minimum \( T_{\text{min}} \) temperature, daily mean relative humidity, minimum relative humidity, and surface air pressure, were taken from the China Meteorological Administration (CMA). To describe the absolute content of moisture in air, we calculated the daily mean specific humidity \( \text{SPH}_{\text{mean}} \) and minimum specific humidity \( \text{SPH}_{\text{min}} \) using the above-mentioned variables. The meteorological data covered the same period as the GIMMS NDVI data. The meteorological sites used in this study were selected from the available CMA sites across the NCP. In this study, the NCP refers to the plains area (lower than 400 m) in the domain of 32.5°N–40°N, 112°E–122°E (figure 1). This study focused only on the cropland area of the NCP.
The selected meteorological sites meet the following criteria: (1) having an elevation lower than 400 m to ensure that the site falls in the plain; (2) being surrounded by cropland and outside of urban areas; and (3) having no missing meteorological observations from March to June in the period of 1982–2006. The land use/cover types surrounding the meteorological sites were identified from the USGS 1 km resolution global land cover characterization (GLCC) dataset.

In the regional climate modelling with the WRF model, the NCEP/DOE Reanalysis II dataset from 1997 to 2006 (dataset number ds091.0 from http://rda.ucar.edu/) was used as the lateral boundary forcing. The NCEP/DOE Reanalysis II has a spatial resolution of 2.5° × 2.5° and a 6 h time interval. The land use data used in the modelling are the default land use data of the WRF model, which are also derived from the USGS 1 km resolution GLCC dataset.

2.2. Data analysis

Spring is defined in our study as the second half of March to the first half of May, and early summer is defined as the second half of May to the second half of June. The winter wheat was mowed in the second half of May. Thus, the spring NDVI was at the first peak, and the early summer NDVI was at the valley of the seasonal NDVI cycle. At each selected site, the averaged NDVI in the 5 pixel × 5 pixel window surrounding the site was calculated. The averaged NDVI value was used to represent the site vegetation greenness. The time series of spring and early summer mean NDVI from 1982 to 2006 were then calculated for each site. The time series of mean climatic variables (i.e., $T_{\text{mean}}$, $T_{\text{max}}$, $T_{\text{min}}$, SPH$_{\text{mean}}$ and SPH$_{\text{min}}$) for spring and early summer were also computed for each site. The changes in the mean values between the time periods of 1982–1991 and 1997–2006 were computed for the climatic variables and NDVI. The spatial correlations between changes in the climatic variables and NDVI were analysed.

2.3. Experimental design

The WRF model version 3.4.1 with the Advanced Research WRF (ARW) dynamics solver (Skamarock et al 2008) was used to diagnose the climatic effects of surface greenness changes. The Noah Land Surface Model (Noah LSM hereafter) was used to simulate land surface meteorological parameters and fluxes. The coupled WRF–Noah model has been used widely to study regional climate changes (e.g., Leung et al 2006) and land–atmosphere interactions (e.g., Zhang et al 2008, 2011, Ge et al 2013). In the Noah LSM, there is one layer of vegetation canopy. The canopy properties are parameterized using green vegetation fraction (GVF), leaf area index (LAI), snow-free albedo (ALB), emissivity coefficients (EMS), etc. The GVF determines the land area directly exposed to the atmosphere and thus the direct evaporation from the shallow top soil layer, $E_{\text{dir}}$. GVF is also one of the parameters determining precipitation interception by the canopy and thus regulates the canopy evaporation, $E_c$. LAI determines the canopy surface resistance and then regulates canopy transpiration, $E_t$. The parameters LAI, ALB and EMS are derived from GVF in the model. The GIMMS NDVI data was used to estimate the GVF following the method of scaled NDVI (Gutman and Ignatov 1998). Although the mean GVF estimates may have biases (e.g., Jiang et al 2006, Jimenez-Munoz et al 2009), the temporal variations of GVF can be reasonably captured (e.g., Zeng et al 2003). The semi-monthly GVF was linearly interpolated to a daily interval and was aggregated to the WRF model grids (20 km × 20 km).

We performed two simulations, which are pre- (PREV) and post-vegetation change (PSTV) experiments. Semi-monthly GVFs were set as the 1982–1991 and 1997–2006 mean values in the PREV and PSTV experiments, respectively. The revisions were implemented only for the 1172 cropland cells of the NCP (figure 1). Excluding the GVF and associated LAI, ALB, EMS (see supplementary figure S1 available at stacks.iop.org/ERL/8/024038/mmedia), PREV and PSTV used the same settings and boundary conditions.
Figure 2. Regional mean semi-monthly NDVI for the 1172 cropland cells in the North China Plain (cropland cells in the domain of 32.5°N–40°N, 112°E–122°E, see figure 1).

We used two nested grids with spatial resolutions (grid numbers) of 60 km (79×89) and 20 km (118×121), respectively. The model domains covering the NCP were centred at 36°N, 117°E (figure 1). The vertical grid contains 28 sigma levels from surface to 100 hPa. The main physical parameterizations used by WRF are the WSM 3-class simple ice microphysics scheme (Hong et al 2004), the Grell–Devenyi ensemble convective parameterization (Grell and Devenyi 2002), the Community Atmospheric Model (CAM3) radiation package (Collins et al 2004), the Yonsei University planetary boundary layer scheme (Hong and Pan 1996), and the Noah land surface scheme (Chen and Dudhia 2001).

The PSTV and PREV experiments were driven by the lateral boundary forcing from 1 January to 31 July of each year in the 10 year period 1997–2006. January and February were used as spin-up simulations. The 10 year ensemble means of the PSTV and PREV experiments from March to July were calculated and compared.

3. Results

3.1. Observed changes in NDVI and climate

Figure 2 shows the seasonal cycle of mean NDVI of the 1172 cropland cells of NCP in 1982–1991 and 1997–2006. Comparing 1997–2006 to 1982–1991, we found greening surfaces in the spring and browning surfaces in the early summer. The NDVI in 1997–2006 was much greater in the spring and smaller in the early summer than in 1982–1991. The spring NDVI increase was 0.04 (from 0.36 to 0.40) and the early summer decrease was 0.02 (from 0.40 to 0.38). The largest increase (0.05) occurred in the first half of April and the largest decrease (0.05) occurred in the second half of June.

Figures 3(a) and (d) illustrate the spatial variability of NDVI changes from 1982–1991 to 1997–2006 in NCP cropland. In the spring (figure 3(a)), the NDVI increases in the central area of NCP were much larger than in the northern and southern NCP regions. The largest NDVI increase (0.15) occurred at the Chaocheng site (36.23N, 115.67E) in the central area of the NCP. The northern and southern NCP regions did not show detectable increases in NDVI. In early summer (figure 3(d)), the NDVI decreases were much larger in the southern and northern NCP than in the central NCP. The largest NDVI decrease (0.1) occurred at the Gaoyou site (32.8N, 119.5E) in the southern NCP. The NDVI changes in the central NCP were small.

Figures 3(b) and (e) illustrate the spatial variability of climate warming from 1982–1991 to 1997–2006 in NCP cropland. In the spring (figure 3(b)), the central NCP experienced less warming than the southern and northern NCP. The Dezhou site (37.43N, 116.32E) in the central NCP cooled by 0.15°C, while the largest warming (1.83°C) occurred at the Zhumadian site (33.00N, 114.01E) in the southern NCP. Climate warming was also larger in the southern and northern NCP than in...
the central areas. The Dezhou site (37.43N, 116.32E) in the central NCP cooled by 0.13 °C while the Tangshan site (39.67N, 118.15E) in the northern NCP experienced the greatest warming (1.44 °C).

By comparing figures 3(a)–(b) and (d)–(e), we found that the spatial patterns of temperature changes are generally coherent with those of NDVI changes. The areas with large (small) greening experienced small (large) spring warming, while the areas with large (small) browning experienced large (small) early summer warming. Figure 4 shows the site-based correlations between changes in NDVI and in temperature. It is notable that there are significant ($p < 0.05$) negative correlations between the changes in NDVI and in temperature. Among the temperature variables, the changes in $T_{\text{max}}$ and $T_{\text{min}}$ had the strongest and weakest correlation with the change in NDVI, respectively. Moreover, the correlations in the spring were stronger than those in the early summer. The correlation coefficient of NDVI with $T_{\text{max}}$ is $-0.69$ ($p < 0.001$) in the spring and $-0.66$ ($p < 0.001$) in the early summer, accounting for $\sim 47\%$ and $\sim 44\%$ of the spatial variance of the $T_{\text{max}}$ change, respectively. NDVI explained $\sim 40\%$ of the spatial variance of the $T_{\text{mean}}$ change in the spring and $\sim 26\%$ in the early summer.

Meanwhile, we also found significant ($p < 0.05$) positive correlations between the changes in NDVI and in air humidity. As shown in figure 3(c), in the spring the central NCP experienced an increase in humidity, while the southern and northern NCP experienced decreases in humidity. In early summer (figure 3(f)), a small decrease in humidity occurred in the central NCP, while large decreases occurred in the southern and northern NCP. Figures 4(d), (e), (i) and (j) show the positive spatial correlations between the changes in humidity and in NDVI. $\text{SPH}_{\text{min}}$ had stronger correlations with NDVI than $\text{SPH}_{\text{mean}}$, and the correlations in the spring were stronger than those in the early summer. The correlation coefficient of the relationship between NDVI and $\text{SPH}_{\text{min}}$ was 0.69 ($p < 0.001$) in the spring and 0.44 ($p < 0.05$) in the early summer, accounting for $\sim 48\%$ and $\sim 19\%$ of the $\text{SPH}_{\text{min}}$ spatial variance in the spring and early summer, respectively. For $\text{SPH}_{\text{mean}}$ the correlation coefficients were 0.53 ($p < 0.01$) in the spring and 0.41 ($p < 0.05$) in the early summer, accounting for $\sim 28\%$ and $\sim 17\%$ of the spatial variance, respectively.

3.2. Simulated effects of vegetation increase/decrease

Figure 5 shows the correlations of the changes in GVF with changes in simulated surface air temperature at 2 m ($T_2$), and water vapour mixing ratio at 2 m ($Q_2$) at the 1172 cropland cells of NCP. As in the observational results above, the negative correlations between GVF and $T_2$ and the positive correlations between GVF and $Q_2$ were well captured by the simulations, and the simulated correlations in the spring were stronger than those in the early summer. These correlations demonstrated that the coupled WRF/Noah model captured the cooling–wetting (warming–drying) effects of increased (decreased) surface greenness in NCP cropland.

Figure 6 shows the semi-monthly regional mean differences in $T_2$ and the surface energy budgets of the 1172 NCP cropland cells. We observed cooling in the spring and warming in the early summer. The ensemble mean maximum cooling ($-0.19$ °C) and maximum warming
Figure 5. Scatter plots of cell-based changes (PSTV minus PREV) in green vegetation fraction (GVF) and changes in simulated surface air temperature ($T_2$) and surface water vapour mixing ratio ($Q_2$) for spring (left column) and early summer (right column), for the 1172 cropland cells of the North China Plain, from the ensemble mean of 10 members. The text at the bottom of each plot are the squares of the correlation coefficients and the confidence levels.

(0.22°C) occurred in the first half of April and the second half of June, respectively, corresponding to the maximum increase and maximum decrease in NDVI. These cooling/warming effects could be explained by changes in the Bowen ratio (figure 6(d)). In the spring, the Bowen ratio decreased, whereas in the early summer, the Bowen ratio increased. As we can see, the surface net shortwave radiation ($Q_S$) and surface net longwave radiation ($Q_L$) had no detected changes (figures 6(b) and (c)). The Bowen ratio changes indicated the inverse changes in sensible heat flux ($F_{Hs}$) and in latent heat flux ($F_{Es}$), as shown by figures 6(e) and (f). In the spring, $F_{Hs}$ decreased and $F_{Es}$ increased, while in the early summer, $F_{Hs}$ increased and $F_{Es}$ decreased. Figure 7 illustrates that the increase/decrease in $F_{Es}$ mainly resulted from changes in canopy transpiration, $E_t$. Because the daytime $E_t$ value is much larger than that at night, the effects on meteorological conditions in the day would be larger than those at night. Such findings could explain the observations of stronger daytime correlations of NDVI with $T_{max}$ and SPH$_{min}$ compared to $T_{mean}$ and SPH$_{mean}$. Although ALB and EMS were also not constants, changes in the surface net shortwave radiation ($Q_S$) and surface net longwave radiation ($Q_L$) were too small to be detected (figures 6(b) and (c)) and thus contributed little to changes in $T_2$.

4. Conclusion and discussion

Using meteorological observations and satellite NDVI data, we investigated the change in cropland greenness in the NCP and the effects of that change on surface air temperature. The results showed that while spring cropland greenness increased (Piao et al 2003, Peng et al 2011), early summer cropland greenness substantially decreased from 1982–1991 to 1997–2006. In contrast to the cooling effect of cropland greening in spring (e.g., Bounoua et al 2000, Bonan 2001, Jeong et al 2009a, 2009b), the cropland browning in early summer had warming and drying effects.

The cooling/warming effects of cropland greenness change mainly resulted from the corresponding changes in the partitioning of surface energy between sensible heat and latent heat fluxes. The increased (decreased) greenness corresponded to high (low) canopy transpiration rate and thus more (less) latent heat flux and less (more) sensible heat flux, i.e., a low (high) Bowen ratio. These results indicate that the cooling effects of vegetation transpiration are stronger than the warming effects of surface albedo and emissivity caused by the greening surface in spring and early summer in the NCP (Bounoua et al 2000, Jeong et al 2009a).

This study not only confirmed a cooling effect of surface greening in spring reported by existing studies (e.g., Jeong et al 2009a, 2009b), but also found a warming and drying effect of cropland browning in early summer. These findings have important implications for understanding the regional climate responses to land use changes.
Figure 6. Differences (PSTV minus PREV) in simulated regional mean surface air temperature ($T_2$), surface net shortwave radiation ($Q_S$), surface net longwave radiation ($Q_L$), surface Bowen ratio, surface latent heat flux ($F_{Es}$), and surface sensible heat flux ($F_{Hs}$) for the 1172 cropland cells of the North China Plain from the ensemble mean of 10 members (short red lines denote the median of 10 members).

Figure 7. Same as figure 6 but for (a) direct evaporation from the shallow top soil layer ($E_{dir}$), (b) canopy transpiration ($E_t$), and (c) evaporation ($E_c$).

et al 2009a, 2009b), but also demonstrated a warming effect of surface browning in early summer. These findings help to understand the spatial variation of climate warming amplitude in spring and early summer in the NCP. The importance of the cooling effect of canopy transpiration was also highlighted in this study. Therefore, even beyond the realm of radiative forcing, which includes the effects of surface albedo (Forster et al 2007), the effects of transpiration should also be considered in order to understand the comprehensive geophysical effects of surface changes.

More importantly, our findings showed that even when land use/cover type remains unchanged, variations in surface greenness might still have detectable climatic effects. Such a conclusion implies that we need to consider not only conversions of land use/cover types but also variations in surface greenness to understand climate change. The cooling/warming effects of surface greenness change, similar to the urban heat island effect, may introduce artificial temperature change due to measurements in certain areas and is informative to studies on regional and global temperature change.

Finally, although the observational findings agree well with the numerical simulations, several relevant issues warrant discussion. Although NDVI is a good indicator of surface greenness and has been used extensively, NDVI is influenced by other environmental factors such as surface soil moisture (e.g., Gitelson et al 2002). Therefore, the correlations illustrated by figure 4 may include effects of
surface soil moisture on local temperature. To confirm our findings, it would be valuable to conduct similar studies using other vegetation index datasets. Additionally, this study only analysed the geophysical processes by which vegetation regulates local/regional climate warming. A potential research direction would be to study biogeochemical effects of vegetation variations because vegetation may have the potential to modify local CO₂ concentrations and regulate climate warming.

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