Feedback between surface air temperature and atmospheric circulation in high-temperature weather in East China: a diurnal perspective

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Abstract

This study proposes the generality of surface air temperature (SAT)–atmospheric circulation feedback during high-temperature weather in late July 2003 over East China by using the Advanced Research Weather Research and Forecasting model (WRF; Version 3) simulations with a succession of 24-h integrations, i.e. on a daily scale, the SAT increase leads to a weakened ridge of the western Pacific subtropical high in the lower troposphere (i.e. negative feedback), whereas it leads to a strengthened ridge in the upper troposphere (i.e. positive feedback) and vice versa. Additionally, using the balance equation of temperature, the feedbacks are clarified from the diurnal-variation perspective. This shows many complex details, e.g. the changes in geopotential heights are more complex than those in air temperatures, and the overall daily feedback appears to be dominated by the feedback during the phase with intense daytime surface heating. All of the WRF-modified land surface conditions can lead to large changes in the maximum, minimum, and average SATs over the mean diurnal scale, with generally larger differences induced by land surface schemes than those induced by initial soil moisture, suggesting that the SAT–circulation feedback can be greatly reduced (or amplified) by different land conditions over the diurnal weather scale and that diurnal variations could substantially contribute to longer timescale climate.

Keywords: feedback; surface air temperature; atmospheric circulation; high-temperature weather; diurnal variation

1. Introduction

Owing to the increase in heatwave events under the background of climate change, which exerts a severe impact on human beings and the environment, there are many investigations focusing on effects of land–atmosphere interaction on the development and maintenance of heatwaves, e.g. low soil moisture was shown to favour the formation and persistence of heatwaves (Ferranti and Viterbo, 2006) and to amplify heatwave events (Fischer et al., 2007; Zhang and Wu, 2011; Zeng et al., 2014, ZWZ14 hereafter). Because the land surface provides the lower boundary of the atmosphere, land surface change can strongly influence the surface air variables [e.g. surface air temperature (SAT); He et al., 2016] through which the circulation of the overlaying atmosphere is modified. The atmospheric circulation, in its turn, impacts the lower atmosphere through various dynamic and thermodynamic processes. For the feedback between the SAT and atmospheric circulation in heatwave events, Fischer et al. (2007) suggested that there is a positive feedback between the SAT and atmospheric circulation, i.e. due to reduced soil moisture in the European 2003 summer heatwave, high SAT produced a heat low at the surface and enhanced the positive height anomalies in the upper troposphere (i.e. the rise of the SAT agreed well with enhanced atmospheric circulation). This soil moisture-induced positive feedback is similar to the facts observed by previous studies (e.g. Oglesby and Erickson, 1989). While for the short-range (18 h) East China high-temperature weather in late July 2003, a negative SAT–circulation feedback was found using different land surface schemes (LSSs) in Version 2.2 of the Advanced Research Weather Research and Forecasting model (WRF), or ARW (Skamarock et al., 2005), i.e. higher surface temperature was induced by the increase in surface sensible heating, and it reduced the geopotential heights in the western Pacific subtropical high (WPSH) in the lower atmosphere (or higher surface temperature resulted in weakened low-level circulation, and vice versa; Zeng et al., 2011). Recently, this negative feedback was further confirmed by the ARW (Version 3.3; Skamarock et al., 2008) 24-h simulations using a specific LSS with various soil wetness; moreover, a positive SAT–circulation feedback was identified in the mid-troposphere by ZWZ14.
The above-mentioned investigations proposed the overall unique feedback over East China in the short range (approximately 24h) using various LSSs and initial values of soil moisture in different versions of the WRF. It is well known that simulated land–atmosphere coupling strength is model-dependent (Koster et al., 2004) and that the land–air feedback also varies over different timescales (Fennessy and Shukla, 1999). Then, two questions naturally arise: (1) Are there any general features of the SAT–circulation feedback in the high-temperature weather over East China? and (2) What mechanisms are responsible for the sensitivity of the simulated feedback induced by land surface perturbation at shorter timescales (e.g. hourly)? The answers to these questions not only help to enhance our understanding of the model and the nature of land–atmosphere interactions over China, but they can also demonstrate the robustness of the feedback at longer timescales than in the previous studies (e.g. Fischer et al., 2007; Zeng et al., 2016).

Therefore, we used the ZWZ14 data and designed more WRF perturbation simulations using various LSSs and soil moisture amounts for the East China high-temperature event, and we investigated the temperature–circulation feedback from a diurnal perspective.

2. Methods and data

2.1. WRF simulations and data

ZWZ14 data using the Noah LSS in ARWv3.3 (Skamarock et al., 2008) are used in this study. Additionally, the model configurations are followed (Table S1, Supporting information). There are four LSSs in the WRF, and only three of them have soil moisture outputs and explicitly account for soil moisture sub-daily variation, i.e. the Noah LSS (hereafter NOAH), the Rapid Update Cycle LSS (RUC), and the Plei–Xiu scheme (PX); therefore, RUC and PX were used in conducting further sensitivity simulations. Some major differences in the LSSs are listed in Table S2, which leads to the differences in the simulations.

Additionally, five groups of simulations are designed here to examine the sensitivity of the hot weather simulation to initial soil moisture (hereafter SMOIS). The simulations, in which SMOIS is the same as the National Center for Environmental Prediction (NCEP), Final (FNL) analysis data are taken as the control (CTL) runs, while relative to CTL, SMOIS is changed by +50, +25, −50, and −25% in the WET50, WET25, DRY50, and DRY25 simulation groups, respectively (Fischer et al., 2007; ZWZ14). Note that in these simulations, in addition to the ZWZ14 simulations using NOAH only, the simulations using RUC and PX are further performed (for PX, only the CTL simulations are conducted due to the similarity between the soil moisture-induced sensitivities in the LSSs), i.e. in total, 110 integrations of 24-h length are used for the analyses here, with the data listed in Table S2.

As illustrated by ZWZ14, the model domain covers East China and the surrounding areas with two-way nesting. Following Zeng et al. (2011) and ZWZ14, for the simulation period we chose the hottest phase of the summer, i.e. late July 2003, when extremely high temperatures occurred over the land of south-eastern China (within area ‘D2’; Figure S1).

For the initial and boundary conditions, the FNL data (except for SMOIS) were applied for the period from 0600 UTC 20 through 0600 UTC 29 (i.e. 1400 LT 20 through 1400 LT 29) July 2003, and ten 24-h integrations that were initialized at 0600 UTC (1400 LT) were conducted using the same suite of model setups, except for the LSS choice (Tables S1 and S2).

2.2. The balance equation of temperature change

This study focuses on the SAT–circulation feedback. Because there are inadequate model outputs for the 2-m level to quantitatively access the relative importance of the physical processes that influence the 2-m SAT simulations, ZWZ14 proposed a method using the analogous temperature relationship between the 2-m temperature and the air temperature at the lowest model level (Z1, hereafter; ~30 m above ground). Here, we also apply the method to evaluate the feedback induced by land surface perturbations.

Derived from the first law of thermodynamics, the air temperature change \( \Delta T_c \) is determined by

\[
\Delta T_c = AD + CN + H_i
\]

where \( AD \), \( CN \), and \( H_i \) denote the terms for the advection, convection, and diabatic effects (units: °C s\(^{-1}\)), respectively. Specifically, \( H_i \) considers the diabatic effects of radiation and the sensible and latent heat exchanges. In particular, because the WPSH controlled the weather during late July 2003, \( CN \) represents the influence due to the combination of the subsidence and stratification in the WPSH, which can be modulated by local weather events with significant vertical motion. \( H_i \) is difficult to directly computed; however, with all of the terms in Equation (1), except \( H_i \), computed by model outputs, \( H_i \) can be calculated accordingly. Therefore, the influence of each term on SAT change can also be accessed, in which the feedback of atmospheric circulation (i.e. advection and convection) to air temperature is included.

3. Results and discussion

Following the previous studies (e.g. ZWZ14), the 10-day means of the simulated results over the D2 area are analysed to examine the feedback in a systematic, climatological manner, and simulated differences in hourly variation and daily average are also focused on in this section.
3.1. Hourly variations

3.1.1. General features of the feedback

As addressed by ZWZ14, the general patterns of the SAT are successfully reproduced by NOAH. Figure 1(a) presents the 10-day mean hourly variations in the area-averaged temperatures at various levels in the LSS simulations. The upper-level temperatures tend to lag slightly in phase from the surface to 300 hPa, which clearly indicates that the air temperatures are affected by surface heating and that this behaviour becomes weaker with height (Figure 1(a)). Figure 1(b) shows that, although affected by surface forcings (including the SAT-induced effect), the geopotential heights at the pressure levels exhibit much more complex diurnal variations than the temperatures. Generally, the heights
The overall daily feedback appears to be dominated by the feedback during 0900–1400LT as the SAT gets increasingly higher (also see Section 3.2). Additionally, slight LSS-induced changes in the temperatures and heights for PX and RUC show similar results compared to NOAH (Figures 1(e)–(h)), which further confirms the above feedback.

As discussed by ZWZ14, the SAT–circulation feedback in the lower atmosphere can be explained by Equation (1). Figure 2 presents the LSS-produced Z1 diurnal variation in the temperature change ($T_c$) and its forcing terms (i.e. AD, CN, and $H_t$). These terms show similar variation in the simulations (Figures 2(a)–(c)), with the most striking feature being the substantial contrast between the forcings during the night-time (∼1900–0600LT in July) and those during the daytime. At night, the $H_t$ cooling is stronger than the CN warming, and as a result of all these terms, $T_c$ has negative values for each LSS, which leads to a cooling night-time SAT (Figure 1(a)). At ∼0700LT, there is an abrupt $H_t$ increase because of the night-time to daytime transition, and the diabatic processes have a clear warming, i.e. apparently different from the night-time cooling. During daytime, the subsidence has a slightly weaker effect on $T_c$ than it does at night largely due to the approximately neutral stratifications in the hot, sunny days of late July, and the diabatic $H_t$ term substantially dominates the CN subsidence term; therefore, the CN – $H_t$ amplitudes (differences) during the daytime (e.g. during 0800–1000LT) can be much larger than those during the night-time. In short, in addition to affecting temperature advection, atmospheric circulation (e.g.

![Figure 2. The diurnal variations in the area-averaged temperature change terms and in their differences at Z1 in the three LSS simulations, where AD, CN, and $H_t$ represent the terms in Equation (1) for local temperature change ($T_c$), by the effects of advection, convection, and diabatic processes, respectively; i.e. $T_c = \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T + \mathbf{v} \cdot \nabla T$, CN = $w(\gamma_x - \gamma)$, and $H_t = T_x - AD - CN$, where $u$ and $v$ represent the horizontal wind speeds in the $x$ and $y$ directions, respectively, $w$ is the vertical velocity, and $\gamma_x$ and $\gamma$ are the adiabatic lapse rate and the temperature lapse rate, respectively, and therefore all these terms can be calculated from the model outputs (ZWZ14).](image-url)
subidence) alleviates the SAT cooling at night and enhances the SAT rising during the daytime.

### 3.1.2. Sensitivity induced by LSS perturbation

Large LSS-induced differences can still be clearly identified. For instance, there exist quite large PX – NOAH differences in the SAT and the 700-hPa height that amount to 0.8 °C and 1.1 gpm at 1400 LT, respectively. The PX temperatures are lower than those of NOAH, resulting in generally lower heights (Figures 1(g) and (h)). Similar results can also be observed from the difference fields, e.g. by moving from NOAH to PX for 1400 LT, SAT is decreased by over 2 °C, corresponding to a surface pressure increase of ~0.5 hPa and a 300-hPa height decrease of ~1.0 gpm over a large area of the upper domain (Figures S2(b), (f), and (j)). All of these results suggest that the LSS choice can greatly modify the SAT–circulation feedback both in amplitude and in overall bias.

Moreover, although the daytime forcings are little changed by the LSSs, the night-time terms are greatly affected (Figure 2). Among the three LSSs, PX gives the largest CN – H difference while RUC presents the least. When moving from PX to RUC (Figures 2(b) and (c)), CN, which is larger than AD in PX, decreases in RUC, suggesting that the relative importance of the factors in the feedback can be reversed by the LSS choice.

#### 3.1.3. Sensitivity induced by soil moisture perturbation

In addition, some land perturbation of soil variables (e.g. soil moisture) also modify the SAT–circulation feedbacks. Compared with the LSS-induced differences (Figure 1), SMOIS-induced differences appear much more regular: when the soil moisture decreases, the SAT and the 850-hPa temperature basically increase for a specific LSS, while at the other levels, temperatures remain almost invariant. The substantial changes appear during ~1100–1400 LT when daytime surface heating is intense, and even a slight change in upper-level temperatures can be identified (Figures 3(a), (c), (e), and (g)).

Correspondingly, compared with the LSS-induced differences, SMOIS-induced SAT height changes are also quite regular. There is a clearly negative SAT–circulation feedback in the lower (i.e. 850 hPa or below) troposphere together with a positive feedback in the upper (i.e. 700 hPa or above) layer. Additionally, NOAH 850-hPa height changes appear larger than the RUC simulations (e.g. Figures 3(b) vs 3(f)), suggesting that SMOIS-induced sensitivity on the feedback is LSS-dependent. In addition, daytime surface heating is the major forcing compared to the night-time, and lower soil moisture tends to present higher SATs particularly during daytime (e.g. Figure 4). Meanwhile, large differences can be seen, e.g. as compared to CTL after 24-h integrations, DRY50 has an SAT increase of ~1.7 °C, corresponding to an 850-hPa height decrease of 2.8 gpm and a 500-hPa height increase of 1.5 gpm over the domain (e.g. Figures 3(c) and (d)).

### 3.2. Daily averages

For the overall comparison, Table 1 lists the results of the 10-day means of the SAT, surface pressure, and geopotential heights using the hourly values during the 10 days. The comparison among NOAH, PX, and RUC clearly shows the existence of a negative (positive) feedback in the lower (upper) troposphere, e.g. the RUC – PX differences in the mean SAT, surface pressure, and 500-hPa height (i.e. RUC – PX differences in $T_{\text{avg}}$, $P_{\text{surf}}$, and $H_{50}$ in the table) are 1.5 °C, −11.7 Pa, and 0.5 gpm, respectively. In other words, the present work confirms the negative SAT–circulation feedback in the lower troposphere (Zeng et al., 2011) and further indicates the LSS-induced positive feedback in the upper troposphere (which is similar to the SMOIS-induced feedback by ZWZ14) based on the overall 24-h means.

Because in climate simulation/projections, the statistics of climatic extreme events are generally based on statistical parameters such as the maximum, minimum, and average SATs (e.g. Im et al., 2008; Zhang and Wu, 2011) on a daily timescale, and because SMOIS would most benefit climate forecasting for some regions (Guo et al., 2011) and could have an impact as large as the LSSs in the simulations (Zeng et al., 2015), the differences in the parameters (e.g. the RUX – PX differences in Table 1) are suggested to affect simulated climates in addition to the high-temperature weather simulations.

The above-mentioned feedback obtained by daily averages is consistent with the daytime results. Similar results were also reported previously in a warm-season convection event (Trier et al., 2008), e.g. initial soil wetness exerts an impact on the atmosphere, especially during the phase of intense daytime surface heating. This is largely because of the large daytime surface net radiation (largely induced by solar radiation), which is mainly partitioned into sensible and latent heat fluxes and presents a relatively strong land–atmosphere interaction.

Although the negative SAT–circulation feedback in the lower atmosphere in this work is different from the positive feedback in the 2003 European heatwave (Fischer et al., 2007), both feedbacks do not contradict to each other in terms of their physical basis. For instance, in the 2003 European heatwave, the low moisture induced higher-than-normal SAT and further generated (strengthened) a weak thermal low at the surface, i.e. there was a positive feedback in the lower atmosphere (Fischer et al., 2007). However, this kind of weak thermal low, whose effect slightly weakens the strong and persistent WPSH, could not be ‘explicitly’ induced over East China; hence, a negative feedback exists in the lower atmosphere. In the upper atmosphere, higher temperatures correspond to enhanced geopotential heights and vice versa (positive feedback), which is consistent with previous studies (Fischer et al., 2007;
Figure 3. Similar to Figure 1 but for differences induced by initial soil moisture.

Zeng et al., 2011; ZWZ14) and can also be explained by the thermal wind theory.

4. Concluding remarks

This study proposes the generalized SAT–circulation feedback using the same configuration of ARWv3 with both different LSS choice and SMOISs. Over daily timescales, the SAT warming, which is induced by land perturbations, results in a weakened ridge of the WPSH in the lower troposphere (negative feedback) and a strengthened ridge in the upper troposphere (positive feedback) and vice versa.

Unlike previous studies, this work focuses on the feedback from the diurnal variation perspective. The feedback shows many complex details in the diurnal variations, in which geopotential heights change much more complexly than air temperatures. Additionally, the overall daily feedback is suggested to be dominated by the feedback during the phase with intense daytime surface heating, with weak feedback of a different nature over some other time periods.

Using the diagnostics of the temperature equation for the surface layer, the feedback can be further explained. There is a substantial contrast between the forcings (e.g. the temperature-induced diabatic processes and the adiabatic process of circulation-induced
of the simulations (unit: gpm). Geopotential heights minus 1400, 3000, and 5500 gpm at 850, 700, and 500 hPa, respectively) in the 10-day mean hourly variations on behalf of the Royal Meteorological Society.

Table 1. Maxima \( T_{\text{max}} \), minima \( T_{\text{min}} \), averages \( T_{\text{avg}} \), 24-h changes \( T_{\text{chg}} \) of 2-m SAT (unit: °C), 24-h averaged index of surface air pressure (i.e. \( P_{\text{sat}} \), the simulation value of surface pressure minus 98000 Pa, for presentation purposes; unit: Pa), and 24-h averaged indices of geopotential heights at 850, 700, and 500 hPa (i.e. \( H_{850}, H_{700}, \) and \( H_{500} \), which are equal to the simulation values of geopotential heights minus 1400, 3000, and 5500 gpm at 850, 700, and 500 hPa, respectively) in the 10-day mean hourly variations of the simulations (unit: gpm).

|                | NOAH DRY50–CTL difference | NOAH DRY25–CTL difference | NOAH WET50–CTL difference | NOAH WET25–CTL difference | RUC DRY50–CTL difference | RUC DRY25–CTL difference | RUC WET50–CTL difference | RUC WET25–CTL difference |
|----------------|---------------------------|---------------------------|---------------------------|---------------------------|--------------------------|--------------------------|--------------------------|--------------------------|
| \( T_{\text{max}} \) | 36.1                      | 34.6                      | 34.1                      | 33.9                      | 33.9                     | 43.5                     | 41.4                     | 38                      |
| \( T_{\text{min}} \) | 27.1                      | 26.9                      | 26.7                      | 26.5                      | 26.4                     | 27.0                     | 26.8                     | 26.4                      |
| \( T_{\text{avg}} \) | 31.4                      | 30.8                      | 30.3                      | 29.9                      | 29.7                     | 33.5                     | 32.6                     | 31.5                      |
| \( T_{\text{chg}} \) | 2.1                       | 0.7                       | –0.2                      | –1.0                      | –1.4                     | 9.6                      | 7.5                      | 4.1                      |
| \( P_{\text{sat}} \) | 494.3                     | 518.3                     | 534.0                     | 543.7                     | 550.2                    | 506.3                    | 519.6                    | 532.7                    |
| \( H_{850} \) | 81.6                      | 82.3                      | 82.6                      | 82.8                      | 82.8                     | 82.1                     | 82.4                     | 82.7                      |
| \( H_{700} \) | 150.0                     | 149.5                     | 149.3                     | 149.1                     | 149.0                    | 149.8                    | 149.6                    | 149.3                    |
| \( H_{500} \) | 402.2                     | 402.0                     | 401.8                     | 401.7                     | 401.6                    | 402.1                    | 402.0                    | 402.0                    |

Figure 4. Similar to Figure 2 but for NOAH differences induced by initial soil moisture.

Diurnal perspective of the modelled temperature–circulation feedback

subidence) during the night-time and those during the daytime, in which subsidence alleviates the night-time SAT cooling while enhancing the daytime SAT rising. Both the LSS and SMOIS choices greatly modify the SAT–circulation feedback in amplitude and overall bias, and even the relative role of the factors in affecting the feedback can be qualitatively reversed by the choices.

We have focused on the ‘local’ aspect (which is associated with the vertical dynamic and thermodynamic processes over the region of interest, e.g. CN and \( H_{I} \)) of the feedback induced by the perturbation, while the ‘remote’ aspect (which is associated with the effect induced by horizontal change, e.g. horizontal circulation can work on SAT variation via AD) is less emphasized. In fact, both the local and remote aspects contribute to the SAT–circulation feedback induced by the perturbation, e.g. CN is larger than AD in the PX simulation (Figure 2(b)), whereas they are comparable in the NOAH and RUC simulations (Figures 2(a) and (c)).

However, detection of the exact mechanism of how the LSSs induced differences in the simulations using the same dataset is a complex issue. First, these differences are generally induced by different parameterizations in physics. For instance, Zheng et al. (2012) found that because of the incorrect formulations of roughness length, a large and cold bias in surface skin temperature over the arid western continental United States was produced. Second, these differences are also due to
the LSS configurations, e.g. the layering of the LSSs is very important (Garratt, 1993), which was also found by Zeng et al. (2015) using the WRF model with NOAH, RUC, and PX included. In this study, we focus on the SAT–circulation feedback. Due to the above-stated complexity, the exact mechanism that causes the differences is beyond the scope of this study.

Finally, because the examination of climatic extremes in climate simulations has been based on the diurnal extreme values (e.g. Im et al., 2008), the statistical SAT means in this work suggest that the differences induced by the perturbation greatly affect the high-temperature weather simulations; furthermore, they probably imply the significance of the SAT–circulation feedback over longer climate timescales.

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Supporting information

The following supporting information is available:

Table S1. Data and physics schemes used in the tests, where for the perturbed SMOIS tests, the values were modified following Zeng et al. (2014) (for details of the physics schemes, see Skamarock et al., 2008).

Table S2. Some major differences between the used LSSs (or tests).

Figure S1. The studied large (D1) and nested (D2) domains and the topography (unit: m) within the nested one.

Figure S2. The 10-day mean difference fields of (a–d) the Z1 SAT (°C), (e–h) surface air pressure (Pa), and (i–l) 300-hPa geopotential height (gpm) induced by the differences in LSS and timing.

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