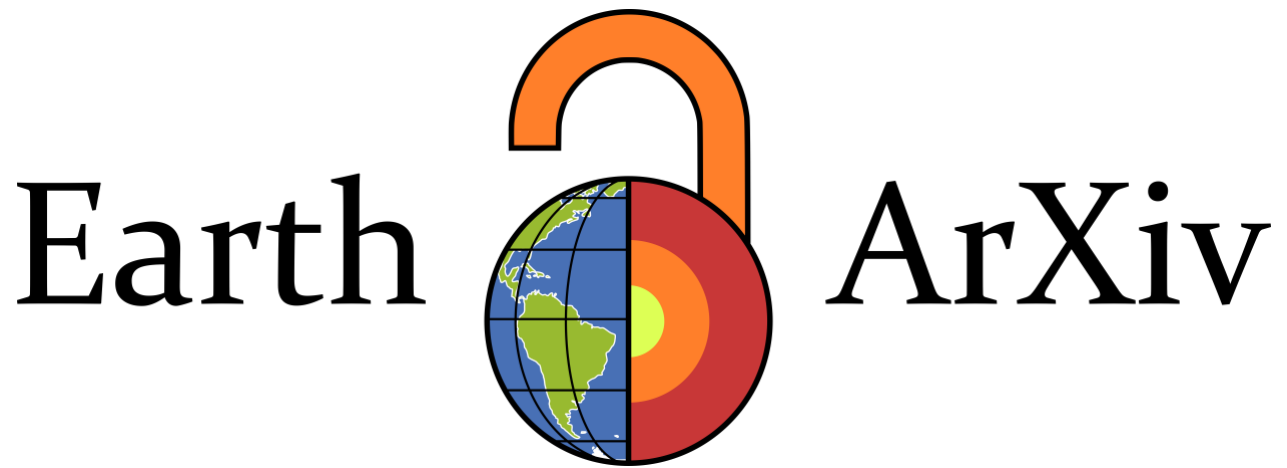




This is a non-peer-reviewed preprint submitted to EarthArXiv.

**Hotspots of Efficient Large-scale Circulation Response to Idealized
Diabatic Heating**

SeungUk Kim^a, Zhuo Wang^a, Francina Dominguez^a

^a *Department of Climate, Meteorology, and Atmospheric Sciences, University of Illinois at Urbana-Champaign,
Urbana, Illinois*

Corresponding author: Francina Dominguez, francina@illinois.edu

Preprint

ABSTRACT

Soil moisture anomalies can influence large-scale atmospheric circulation and thereby affect remote regions. However, the seasonal and geographical dependence of these circulation responses has been largely unexplored. Here, we conduct idealized experiments with a Linear Baroclinic Model, imposing identical low-level diabatic heating anomalies across global land to examine atmospheric circulation responses to persistent drought-like heating. Responses are generally strongest in winter, weakest in summer and are characterized by an upper-level anticyclonic anomaly displaced northeastward from the heating source, with the largest displacement in spring.

Several hotspots emerge from our analysis: heating over the Great Plains, Tibetan Plateau, southeastern South America, and Australia produces substantially stronger circulation responses than heating elsewhere. These stronger responses are associated with enhanced vertical motion, reflecting strong low-level convergence and upper-level divergence. The hotspot regions are characterized by lower climatological static stability and enhanced warm advection. Warm advection is strongest when the imposed heating is located near elevated terrain and the upper-level jet, where strong temperature and circulation perturbations interact with intense upper-level westerlies. Perturbed temperature gradients contribute to the northeastward displacement of the upper-level response through thermal wind balance. Our results suggest that sustained heating in these hotspot regions can generate strong atmospheric circulation anomalies, providing a dynamical pathway for drought amplification and spatiotemporal propagation.

SIGNIFICANCE STATEMENT

If you wish to include this optional element, paste the text here. It should be fewer than 120 words. Additional information and a list of eligible journals is available online.

1. Introduction

Soil moisture is a key variable in the climate system, providing a link between land surface and atmosphere (Seneviratne et al., 2010). Studies on land-atmosphere interaction have largely focused on the role of soil moisture in local atmospheric variables. Soil moisture anomalies modulate evaporation and thereby affect the partitioning of net surface shortwave radiation between latent and sensible heat fluxes (Betts et al., 1996; Santanello et al., 2018). The resulting changes in surface heat and moisture exchange can modify local surface temperature, boundary layer development, and precipitation (Koster et al., 2004; Cook et al., 2006; Taylor et al., 2012; Miralles et al., 2012; Guillod et al., 2015). Through these feedbacks, land-atmosphere coupling can contribute to the development or amplification of extreme events such as droughts and heatwaves and sometimes contribute to adjacent region's extreme events (Fischer et al., 2007; Miralles et al., 2019; Schumacher et al., 2022).

Emerging studies have expanded this local perspective by showing that local soil moisture anomalies can also induce large-scale atmospheric circulation responses extending beyond the forcing region (Koster et al., 2014; Koster et al., 2016; Wang et al., 2019; Teng et al., 2019; Martius et al., 2021; Jiang et al., 2024; Sena and Williams, 2026). These studies have used atmosphere and earth system models of varying complexity, using idealized drought-like anomalies in the land surface or atmosphere to force models, and examined the resulting large-scale atmospheric response, especially during summer. For example, Koster et al. (2016) found that dry land surface anomalies and associated idealized diabatic heating anomalies over North America produced similar upper-level wave patterns, including a pronounced high anomaly over west-central North America. Wang et al. (2019), performed experiments over North America and northern Middle East, and showed that the phase locked spatial structure of the atmospheric response depends on regional orography, which modifies surface pressure and the zonal temperature gradient. Using CESM and completely drying out soil moisture over the Great Plains, Teng et al. (2019) found that synoptic eddies are an important source of circumglobal wave responses during summer.

The identified atmospheric wave patterns induced by land atmosphere interaction in these studies are particularly relevant because persistent amplified Rossby waves are frequently associated with summertime extreme events across the Northern Hemisphere (Schubert et al., 2011; Hoskins and Woollings, 2015). Branstator (2002) showed that the mid-latitude jet can act as a waveguide, supporting circulation anomalies to be meridionally trapped and zonally

68 elongated over large distances and thereby forming circumglobal wave responses. Persistent
69 quasi-stationary Rossby waves can in turn produce regional anomalies in temperature and
70 precipitation, which have been linked to several major extreme events in the Northern
71 Hemisphere. For example, Schubert et al. (2014) showed that stationary Rossby waves
72 played an important role in monthly Eurasian temperature and precipitation variability. Teng
73 et al. (2013) found that the probability of US heatwaves can be increased and preceded by
74 subseasonal planetary waves with a wavenumber of 5. Further, several studies demonstrated
75 that amplified quasi-stationary planetary waves, particularly with a wavenumber of 6–8, are
76 associated with simultaneous summer extreme events across separated regions of the
77 Northern Hemisphere (Petoukhov et al., 2013; Coumou et al., 2014; Kornhuber et al., 2017;
78 Teng and Branstator, 2019). For example, Kornhuber et al. (2020) showed that recurrent
79 summertime atmospheric circulation patterns that can induce simultaneous heat extremes
80 across central North America, Europe, and Asia, have wavenumbers 5 and 7. Given that
81 regional or concurrent extreme events are associated with these Rossby wave patterns, it is
82 important to understand how such large-scale wave responses can result from land surface
83 heating.

84 Although previous studies have shown that imposed drought-like land surface anomalies
85 can induce large-scale atmospheric responses, the conditions under which these responses
86 strongest, and the physical processes that control their strength, remain poorly understood.
87 Most previous work has focused only on boreal summer, when land-atmosphere coupling is
88 generally strong (Koster et al., 2004; Miralles et al., 2012). Whether similar organized
89 circulation responses can occur in other seasons have been less explored. Furthermore,
90 previous studies have examined a limited number of forcing regions, leaving the global
91 geographical dependence of wave responses unexplored.

92 In this paper, we investigate the geographical and seasonal dependence of the
93 atmospheric circulation responses to idealized low-level diabatic heating using the Linear
94 Baroclinic Model. We then examine the physical mechanisms that cause some regions to
95 produce stronger atmospheric responses to heating than others. Section 2 introduces the
96 experimental design and method of analysis. In section 3, we investigate the seasonal
97 circulation responses and identify hotspot regions where heating perturbations generate
98 particularly strong large-scale responses. Then, the strength of circulation responses is
99 approximated using linearized equations in section 4. The approximations are then used in

section 5 to investigate the processes responsible for the enhanced responses in the identified hotspot regions. Section 6 discusses the broader implications and limitations of the results.

2. Experiment and Analysis Method

Our objective is to explore the response of the atmosphere to persistent drought and identify regions that exhibit higher sensitivity than others. To this end, we impose identical drought-like conditions at locations across the globe and examine how the atmospheric response varies depending on the forcing locations. This section describes how we designed these experiments using the Linear Baroclinic Model and how we condensed the results for analysis.

a. Linear Baroclinic Model (LBM)

The Linear Baroclinic Model (LBM) is a simple numerical model designed to investigate linear responses of the atmosphere to prescribed forcing (Watanabe and Kimoto 2000; Watanabe and Jin 2004). Suppose that we have a vector X of prognostic variables in the primitive equations. The LBM solves the primitive equations linearized about a basic state $\bar{X}$ with nonlinear terms neglected. For a steady forced problem, the linearized primitive equations can be written as

$$LX' = F'$$

where $L \equiv L(\bar{X})$ is the linear part of the dynamical operator, which is a function of the basic state. Then, the steady-state response of the atmosphere (X') to imposed forcing (F') can be obtained as,

$$X' = L^{-1}F'.$$

The LBM has a coarse resolution compared with the current general circulation models. In this study, we use the dry version of the model at a horizontal resolution of T21 ($\sim 5.6^\circ$) and 20 vertical levels in sigma coordinates. Owing to its coarse resolution and simplified equations in linearized forms, the LBM has a low computational cost. This makes it well suited for systematically examining the response of the atmosphere to various forcings and isolating physical mechanisms (Lu and Lin, 2009; Jenney et al., 2021; Jiang et al., 2024).

The forcing (F') we used to perturb the model is a low-level diabatic heating anomaly (Q) that mimics drought conditions. Increased sensible heat flux over persistently dry land surfaces can produce low-level heating anomalies. Similar approaches have been used in previous studies (Koster et al., 2016; Wang et al., 2019). The magnitude of the imposed heating is 5 K/day at the surface (0.995σ) and decreases rapidly with height in sigma coordinates, i.e. $Q(\sigma) = 5 \exp[-20(\sigma - 0.995)^2]$ (Fig. 1a). The forcing has an elliptical shape in horizontal directions, with its magnitude decreasing linearly outward from the center. In this way, the imposed heating is confined to the grid cell of interest.

For the basic state ($\bar{X}$), which determines the linear dynamics (L), we use the climatology from ERA40 which is provided as a package with LBM codes. Note that LBM was originally released in the early 2000s. Four seasonal-mean basic states are used for December to February (DJF), March to May (MAM), June to August (JJA), and September to November (SON). The experiments using basic state from NCEP-NCAR reanalysis is also conducted and they agree very well with those using ERA40 (not shown).

To obtain the steady-state solution of the linear dynamic systems, $L^{-1}F'$, we chose time integration method over matrix inversion because time integration is computationally less expensive. Each experiment is integrated for 70 days from initialization, and the mean over the final 5 days is taken to represent the steady-state response. The low-level heating experiment is repeated at every grid cell within 55°S and 85°N for each of the four seasons. This domain has 64 longitude grid cells and 25 latitude grid cells, resulting in 1600 experiments per each season (Fig. 1c), and 6400 experiments in total for the four seasons. As we are only interested in land regions, we are left with 495 experiments per season, 1980 experiments for the four seasons. Because the LBM solves the linearized atmospheric response to a localized prescribed forcing, this series of experiments can be viewed as analogous to a Green's function approach.

An example response to a prescribed diabatic heating anomaly over the Great Plains shows the atmospheric wave response (Fig. 1b). At upper levels (200hPa), the spatial patterns of height and zonal wind responses exhibit a wave-train-like pattern extending northeastward from the forcing location. The maximum increased height is observed near central North America, slightly displaced northeast from the center of the imposed forcing. At lower levels (850hPa), by contrast, decreased height is observed more directly over the forcing location. This circulation response can promote enhanced ascent near the forcing location and

descending motion farther to the northeast, potentially creating atmospheric conditions favorable for drought in adjacent areas. The detailed mechanisms underlying this possible drought propagation pathway are beyond the scope of this study, but we will briefly return to this in the discussion section.

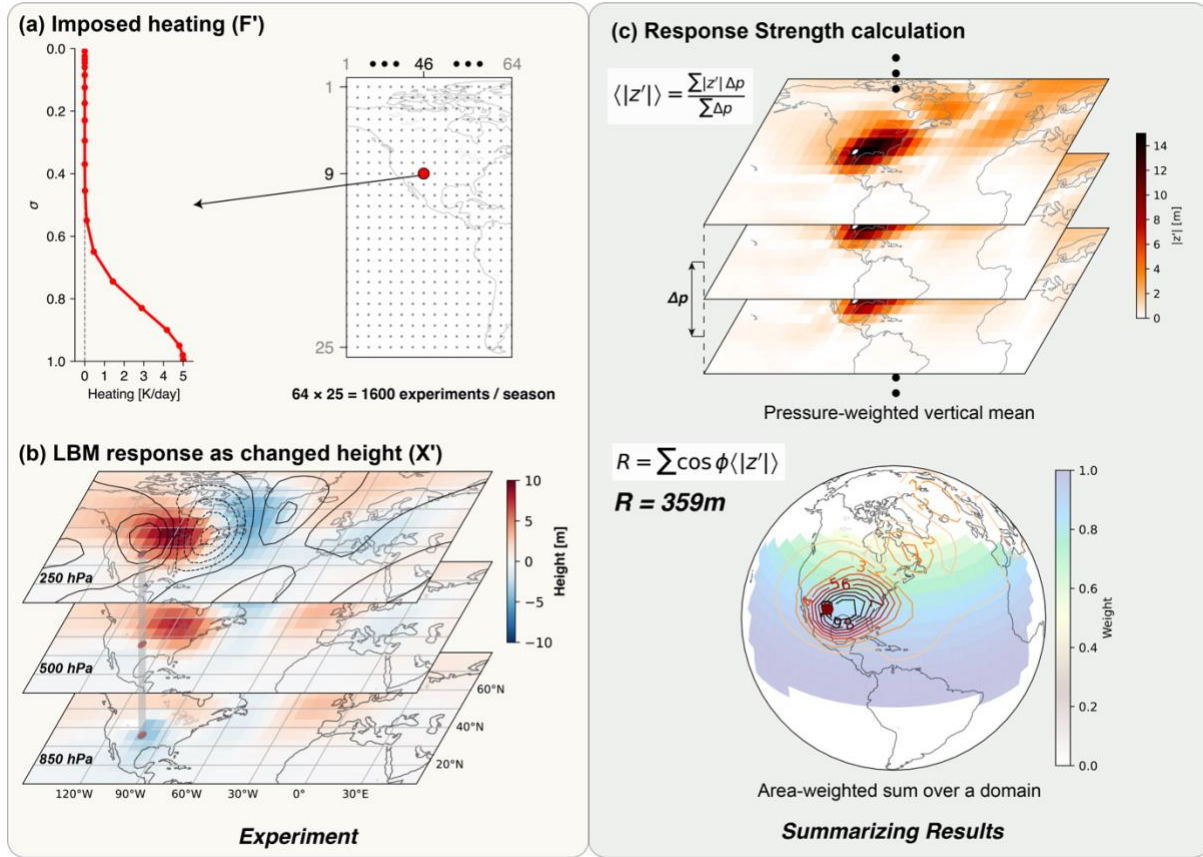


Fig. 1. Experiment design. (a) Heating anomaly used to force LBM and representation of grid cells. (b) Steady-state height response to the forcing applied over the Great Plains at 250, 500, and 850 hPa in MAM season. (c) Demonstration of Response Strength calculation with pressure-weighted mean calculation (top) and the calculated mean drawn on a single level (contour) with applied domain and weight (shading).

b. Response Strength and Topography-aware Regional Mean

The LBM output contains 8 variables at 20 pressure levels, making direct interpretation and comparison across thousands of experiments challenging. To condense and summarize large volume of model output, we introduce a new “Response Strength” (R) metric which represents the overall magnitude of the atmospheric response for each experiment. It is calculated as the area-weighted total geopotential height change ($|z'|$) within a selected domain (Ω), and it is the pressure-weighted averaged across all pressure levels.

$$R = \sum_{(\phi, \lambda) \in \Omega} \cos \phi \langle |z'(\phi, \lambda)| \rangle \quad (1)$$

where $\langle A \rangle = \sum A \Delta p / \sum \Delta p$, ϕ is latitude, and λ is longitude. In this study, the selected domain (Ω) extends 5000km west, 10000km east, 3750km north, and 3750km south of the forcing location, and all grid cells within this domain are included in the calculation (Fig. 1d; shading). This domain roughly captures the extent of significant upper-level wave responses. The asymmetric east-west extent is intended to capture the downstream atmospheric response to the imposed heating.

This metric reduces the 3D results from each LBM experiment to a single value, allowing the overall magnitude of the atmospheric disturbance to be compared across forcing locations. For example, with the same basic state of MAM season, heating imposed over the Great Plains produces a Response Strength of 349m, whereas the same heating imposed over Australia produces 187m. Thus, heating over the Great Plains generates a substantially stronger large-scale atmospheric disturbances than identical heating imposed over Australia.

In order to present 3D responses on 2D maps, we additionally calculate a topography-aware local mean (A_{local}).

$$A_{local} = \left\{ \frac{\sum A \Delta p_*}{\sum \Delta p_*} \right\}_{3 \times 3}$$

where $\{A\}_{3 \times 3}$ represents the area-weighted mean over the surrounding 3 by 3 grid cells, and Δp_* is the layer thickness used in the vertical averaging. The averaging is done only for mid-troposphere below 300hPa and above 850hPa or 50 hPa above surface. This local averaging is intended to capture the atmospheric response in the regions nearby rather than only at the forcing location itself. Although spatial averaging can reduce the magnitude of a strong response of a single grid cell, it provides a more robust representation of the regional signal and aids qualitative interpretation of local variables such as vertical velocity and temperature advection.

c. Multivariate EOF Analysis

To identify the general patterns of the atmospheric response to the diabatic heating anomaly over land, we used empirical orthogonal function (EOF) analysis. Before performing the EOF analysis, we introduce a preprocessing step to better account for differences in the background circulation among forcing locations.

For each forcing location over land grid cells, we construct a domain extending 15000km in the along longitudinal direction and 7500km in the along latitudinal direction, centered on the forcing location, with a grid spacing of 250km. The model outputs of streamfunction (ψ), height (z), zonal (u) and meridional (v) wind are interpolated onto this grid. Grid cells close to the equator located within 5°S and 5°N, whose responses have chances to cross equator are excluded in this analysis.

For each season, the interpolated fields of the four variables are reshaped into a single vector for each experiment. The resulting vectors are then stacked and used in a multivariate EOF analysis. The analysis is performed at each hemisphere, using only the 250hPa responses, where the large-scale wave response is most clearly expressed. This allows us, first, to characterize the mean response across all forcing locations within a season, and then, to identify the dominant modes of variability for each season.

3. Large-scale Circulation Response to Diabatic Heating

a. Response Strength Hotspots

Our analysis reveals several hotspot regions where the same diabatic heating anomaly consistently produces stronger circulation responses than elsewhere (Fig. 2 for the Northern Hemisphere; Fig. S1 for the Southern Hemisphere). These results are based on 495 heating experiments repeated under four seasonal basic states, for a total of 1,980 experiments.

These regions are mostly located along the mid-latitudes, including east of the Rocky Mountains over the US Great Plains, Sahara Desert, and Tibetan Plateau in the Northern Hemisphere (Fig. 2), Southeastern South America, Southern Africa, and Australia in the Southern Hemisphere (Fig. S1). It has been reported in recent studies that imposed heating over the Great Plains (Koster et al., 2016; Wang et al., 2019), Tibetan Plateau (Jiang et al., 2024), and Australia (Martius et al., 2021) can produce circumglobal wave responses through imposed dry surface conditions. We also tested different box sizes and different position of forcing location within the box and found consistent spatial pattern of the response strength map (Fig. S2).

Except for the Sahara region, other regions are mostly located near complex topography. Our results suggest that if there is a persistent heating anomaly over these ‘hotspots’, for

example due to persisting dry land surfaces, there is a higher chance for the atmosphere to be perturbed and generate large-scale circulation anomalies that might affect synoptic weather systems in adjacent regions.

The magnitude of R varies by season with the strongest signals during winter (DJF) and the weakest during summer (JJA). Land grid averaged response strengths were 273.2 m for winter and 164.8 m for summer. The seasonal magnitude fluctuation is well pronounced over hotspot regions where cold season signals can be relatively small compared to warm seasons. A potential cause of seasonally varying magnitude can be related to the basic state of each season (Fig. S3), especially the meridional temperature gradient. Given the year-round warm equatorial temperature, colder land surfaces in the Northern Hemisphere during winter cause stronger meridional temperature gradients and stronger zonal wind in the upper atmosphere. Latitudinal shifts in the location of these hotspots reflect the location of upper-level jet. Note that our experiment uses exactly the same heating, which does not reflect the seasonally varying atmospheric responses to drying. In the real world, different climatological soil moisture and planetary boundary layer profiles would generate different heating responses in the atmosphere to the same soil moisture anomalies in different seasons.

While different basic states can affect the magnitude of the patterns, hotspot regions exhibit stronger R relative to adjacent regions regardless of the basic state. This suggests that such strong responses are not only dependent on the atmospheric state but also on the specific geographical characteristics, such as terrain. The factors affecting the strength of the response will be quantitatively analyzed in detail in the section 4.

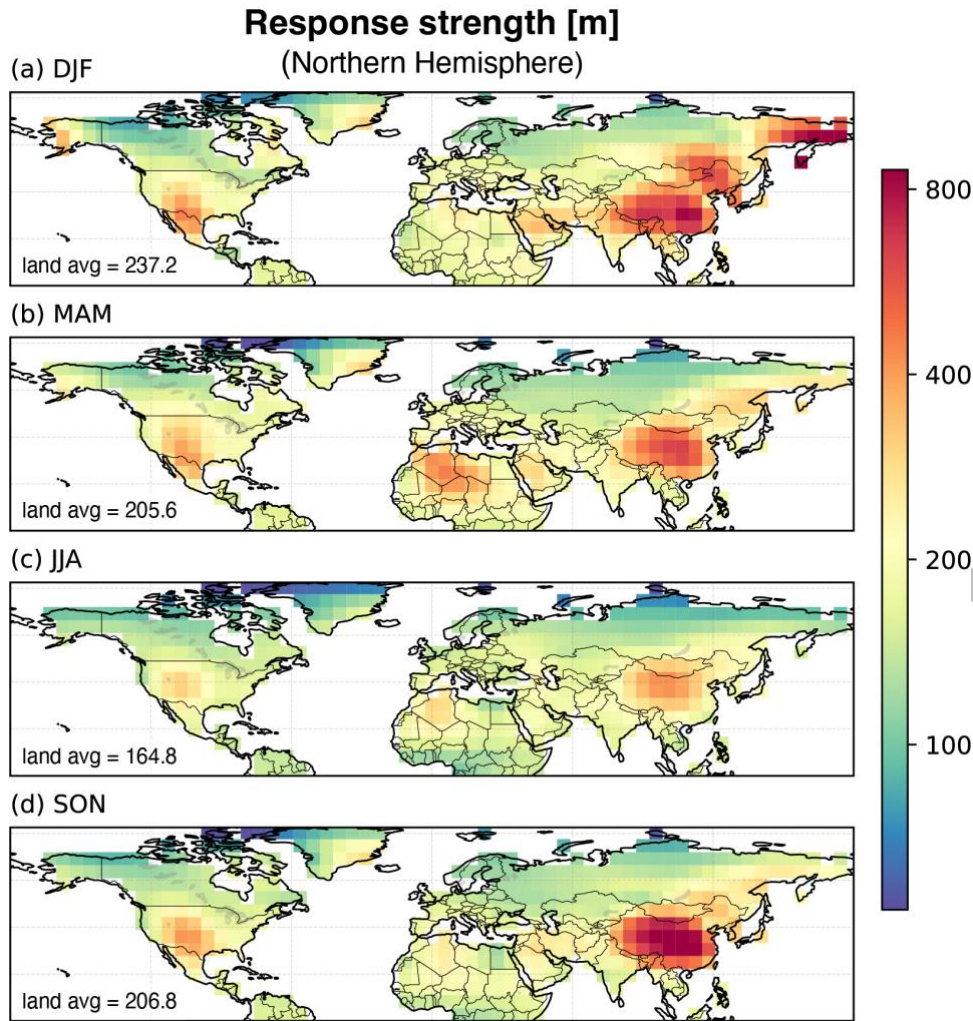


Fig. 2. Seasonal response strength in Northern Hemisphere. Spatial patterns of response strength are shown for each season, with shadings displayed on a log-scale colormap. Land grid cell averages are also shown bottom left of each panel.

b. Induced Large-scale Circulation Changes

Diabatic heating anomalies imposed over land triggers large-scale circulation responses characterized by upper-level wave trains. The mean response and multivariate EOF analysis using the responses at 250hPa interpolated onto 15000km by 7500km domain centered on each forcing location, show characteristic responses of the atmosphere to imposed diabatic heating (Fig. 3). In general, the mean response and EOF1 respectively feature the patterns of the ridge where atmospheric wave starts (Fig. 3 left), the first downstream trough of the wave response (Fig. 3 middle). The mean responses exhibit strong positive streamfunction anomalies (white crosses) near the forcing location, followed by weaker negative anomalies downstream at 250hPa (Fig. 3 left). Except in summer, the first mode features pronounced

negative streamfunction anomalies downstream (red crosses), completing a wave-like patterns (Fig. 3 middle). The geographic distribution of strong principal components well captures the location of hotspot regions (Fig. 3 right), indicating that these EOF patterns can be interpreted as representative circulation responses to low-level diabatic heating imposed over hotspots. Thus, responses in these hotspot regions produce fully developed wave patterns with more well-defined downstream responses.

Although the average R is the greatest in winter (DJF; Fig. 2), the magnitude of the mean response is smaller than that of other seasons. In addition, EOF1 explain only 28.0% of the total variance, substantially less than in the other seasons, indicating greater spatial variability among winter responses. However, winter responses exhibit stronger wave-like signals, with strong downstream propagation captured by EOF1. These patterns are associated with larger absolute values of changed geopotential height ($|z'|$). The center of the positive streamfunction anomaly is displaced east (mean) relative to the forcing location, suggesting that the upper-level wave train starts to form northeast of the imposed heating rather than directly above. Stronger upper-level winds during winter, and associated stronger anomalous vorticity advection, may contribute to these stronger downstream wave responses.

The spring (MAM) response broadly resembles the winter response, but it exhibits weaker downstream signals in the first two dominant modes of EOF. However, the centers of the strongest positive streamfunction anomalies (white crosses) show particularly large northeastward displacements from the forcing location in EOF1. These suggest that imposed heating in spring could substantially change large-scale circulation in the adjacent downstream regions, potentially affecting the region's synoptic weather if the heating is strong enough.

In contrast, the summer (JJA) response exhibits the weakest wave-like structure and the smallest average response strength, potentially reflecting the weaker summertime jet. Although the mean response has strong positive streamfunction anomalies, these anomalies remain concentrated near the forcing location. All downstream negative streamfunction anomalies in mean, EOF1 are weaker than in the other seasons. These features suggest that imposed diabatic heating during summer has a more limited ability to generate strong remote responses related to linear dynamics, while it could be efficient in perturbing the circulation locally. However, because soil moisture anomalies during the warm season, when soil is typically drier, can generally induce stronger sensible heat flux responses, summer soil

moisture anomalies may have a higher chance to generate stronger low-level heating anomalies compared to other seasons, which in turn would generate stronger non-linear circumglobal atmospheric responses as described in Teng et al. (2019).

Similar to the spring season in both its wave structure and magnitude of average R, the fall (SON) response exhibits wave-like patterns in the mean and EOF1. But the direction of wave propagation differs between the two seasons. Whereas the spring response exhibits a more equatorward propagation, the fall response shows a more poleward propagation signal. EOF1 is particularly dominant in fall, explaining 47.2% of the total variance, indicating a comparatively coherent leading pattern of circulation response during this season.

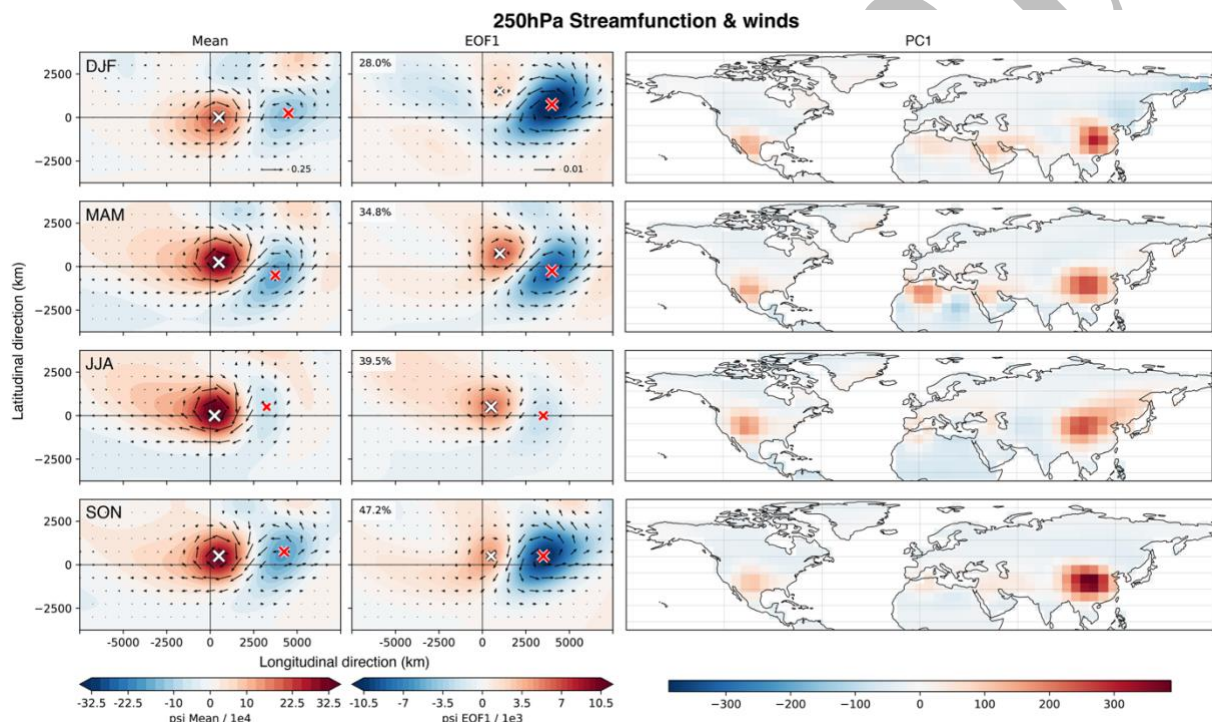


Fig. 3. Induced changes in streamfunction and winds at 250hPa in Northern Hemisphere forcing locations over land as mean, first mode and its principal component of EOF analysis. Left two columns are shown on a 15000km by 7500km grid. The white (red) cross indicates local maximum (minimum). Contour levels are scaled. Explained variances values are shown at the top left corner of each panel. Principal component at each grid cell is depicted on a map.

Example responses at 250, 500, and 850 hPa from four LBM experiments are shown in Fig. 4. It shows different spatial patterns between spring and summer seasons at the Middle East (a region with weak response) and Great Plains (a region of strong response) both located along the same latitude. The spring experiments exhibit clear atmospheric wave

responses, particularly when heating is imposed over the Great Plains, which is one of the hotspot regions (Fig. 2). At both forcing locations, the spring responses at 250hPa and 500hPa strong positive streamfunction anomalies are observed, corresponding to upper-level high pressure system, displaced east or northeast. At 250hPa, these anomalies are followed by alternating downstream anomalies extending southeastward, consistent with the general wave response patterns identified in Fig. 3.

These wave-like features are less pronounced in the summer season responses. Although relatively weak, the summer response in the Great Plains experiment still exhibits signals of propagating waves in the upper level with the strongest perturbation to the north of the forcing center. The structure of observed upper level atmospheric waves in the Great Plains have similar features found in the previous studies with anticyclonic circulation near central North America and northeastward propagation (Koster et al., 2016; Teng et al., 2019). However, the upper-level response in the Middle East experiment shows only faint downstream signals.

At lower levels, the circulation responses have simpler structure and are considerably more localized. Unlike the upper-level response, the location of the strongest negative anomalies shows little displacement from the forcing location regardless of the season. At 850hPa, negative streamfunction anomalies, corresponding to cyclonic circulation, develop around the forcing center and remain collocated with the imposed low-level diabatic heating.

Thus, drought-like heating anomalies induce cyclonic circulation in the lower atmosphere near the imposed heating source, whereas anticyclonic circulation is induced in the upper atmosphere and is displaced to the northeast of the forcing location. The upper-level anomaly can further develop into a downstream wave train, with different strength and spatial extent depending on the season and the location of imposed heating. In particular, the identified ‘hotspots’ are more efficient at generating such large-scale circulation anomalies in response to the same imposed heating.

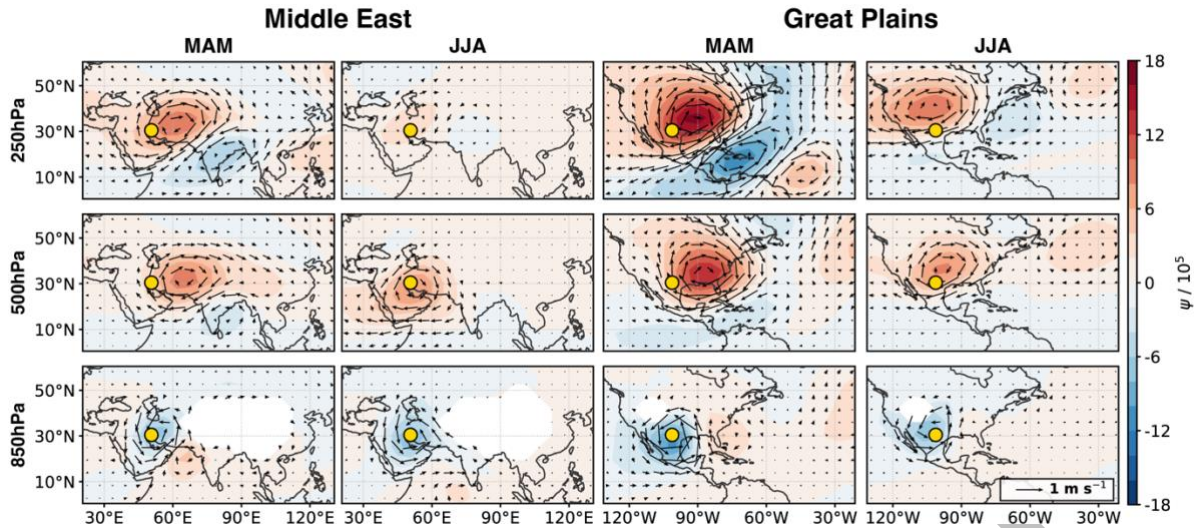


Fig. 4. Sample perturbed streamfunction and wind vector in the Middle East and Great Plains in MAM and JJA at selected levels. Shadings indicate streamfunction. Arrows indicate wind vectors. Reference wind vector is on the bottom right corner. Yellow circle is the location of imposed forcing in each simulation.

4. Rossby Wave Source

The existence of ‘hotspots’ raises a further question: why are these regions more efficient at triggering large-scale atmospheric responses to the same imposed heating? To identify the mechanisms responsible for these stronger responses, we derive an approximated expression for the magnitude of the geopotential height response and perform scale analysis of its individual terms. This framework allows us to assess the relative importance of the physical processes that contribute to stronger wave responses.

We focus on the spring season, when the upper-level positive streamfunction anomalies exhibit the largest northeast displacement from the forcing location, indicating a strong potential to affect circulation over adjacent regions. Also, springtime drought has critical impact on agriculture, as spring is the beginning of the growing season, so that persistent drought-like anomalies are particularly relevant. We therefore use the spring response to illustrate the underlying mechanisms in detail. The main conclusions are broadly consistent across other seasons.

a. LBM height response approximation using RWS

From the vorticity equation, we decompose the wind vector into rotational and divergent wind to derive Rossby Wave Source (Sardeshmukh and Hoskins, 1988).

$$\left(\frac{\partial}{\partial t} + \vec{v} \cdot \nabla\right) \zeta_a = -\zeta_a \nabla \cdot \vec{v} \quad (2)$$

where ζ_a is absolute vorticity, which is the sum of planetary vorticity (f) and relative vorticity (ζ), and $\vec{v}$ is wind vector. The wind vector can be decomposed into rotational and divergent wind defined as

$$v = v_\psi + v_\chi, \quad v_\psi \equiv \hat{k} \times \nabla \psi, \quad v_\chi \equiv \nabla \chi.$$

Plugging this into Eq. 2 and rearranging terms gives

$$\left(\frac{\partial}{\partial t} + \vec{v}_\psi \cdot \nabla\right) \zeta_a = -\vec{v}_\chi \cdot \nabla \zeta_a - \zeta_a \nabla \cdot \vec{v}_\chi = -\nabla \cdot (\zeta_a \vec{v}_\chi) \quad (3),$$

Where the right-hand side is defined as Rossby Wave Source (hereafter RWS; S)

$$S \equiv -\nabla \cdot (\zeta_a \vec{v}_\chi).$$

The RWS quantifies how divergent circulation generates or redistributes vorticity (rotational circulation) through vortex stretching ($-\zeta_a \nabla \cdot \vec{v}_\chi$) and transport across background vorticity gradients ($-\vec{v}_\chi \cdot \nabla \zeta_a$). As our goal is to obtain the approximated form of geopotential height changes ($|z'|$) of steady-state response from LBM, we linearize and assume steady-state in Eq. 3 which gives

$$\vec{v}_\psi \cdot \nabla \zeta' + v'_\psi \cdot \nabla (\bar{\zeta} + f) = S'. \quad (4)$$

We omitted vector notation for simplicity. Assuming the streamfunction to have wave form, and writing relative vorticity with streamfunction, we can write above equation as

$$\psi' \approx \frac{S'}{D} \quad (5)$$

where D is a function of wavelength and basic state rotational winds, which is treated as a constant in this study. For more details about derivation and assumptions, see Appendix. This suggests that streamfunction response (ψ') can be approximated with anomalous RWS (S'). Thus, mechanisms that make anomalous RWS large can result in stronger perturbations of streamfunction. Given that streamfunction can be approximated as $\psi \approx \Phi/f$ in the mid-latitude, where Φ is geopotential, and thus $gz \approx f\psi$, we can approximate the magnitude of perturbed heights ($|z'|$) as

$$|z'| \approx \frac{|f|}{g} \frac{|S'|}{|D|} \quad (6).$$

Given that hotspots are mostly located near mid-latitude regions, we approximate anomalous RWS with assumptions that are suitable for mid-latitude conditions. In the mid-latitudes, planetary vorticity ($f \sim 10^{-4} s^{-1}$) typically is about an order of magnitude larger than climatological relative vorticity ($\bar{\zeta} \sim 10^{-5} s^{-1}$) and perturbed relative vorticity ($|f| > |\bar{\zeta}| > |\zeta'|$). Thus, absolute vorticity (ζ_a), that is the sum of planetary vorticity (f) and relative vorticity ($\zeta = \bar{\zeta} + \zeta'$), can be approximated as planetary vorticity alone ($\zeta_a \sim f$) in the mid-latitudes.

Since anomalous RWS quantifies anomalous convergence of absolute vorticity flux that generates rotational circulation (Eq. 3), the above scale analysis reduces anomalous RWS to the contribution from planetary vorticity flux (Fig. 5a) and allow us to neglect the role of climatological (Fig. 5b) and perturbed (Fig. 5c) relative vorticity fluxes. Decomposing the contribution from planetary vorticity flux gives the advection (Fig. 5d) and stretching (Fig. 5e) terms, where the stretching term turns out to be the dominant term to explain total RWS (Fig. 5f). Therefore,

$$S' \approx -f \nabla \cdot \vec{v}'_{\chi}. \quad (7)$$

In other words, anomalous divergent circulation that interacts with background planetary vorticity serves as the dominant source of rotational circulation in the mid-latitude when low-level diabatic heating anomaly is imposed. While this approximation does not hold in tropical regions, where advection of vorticity plays an important role (Sardeshmukh and Hoskins, 1988), the regions highlighted in our analysis are located in the midlatitudes, where this approximation is more appropriate. A similar approximation is valid for other seasons (Fig. S4–6). This gives the first-order approximation of the anomalous RWS and the magnitude of perturbed geopotential height as

$$|z'| \approx \frac{|f|}{g} \frac{|f \nabla \cdot \vec{v}'_{\chi}|}{D} \quad (8)$$

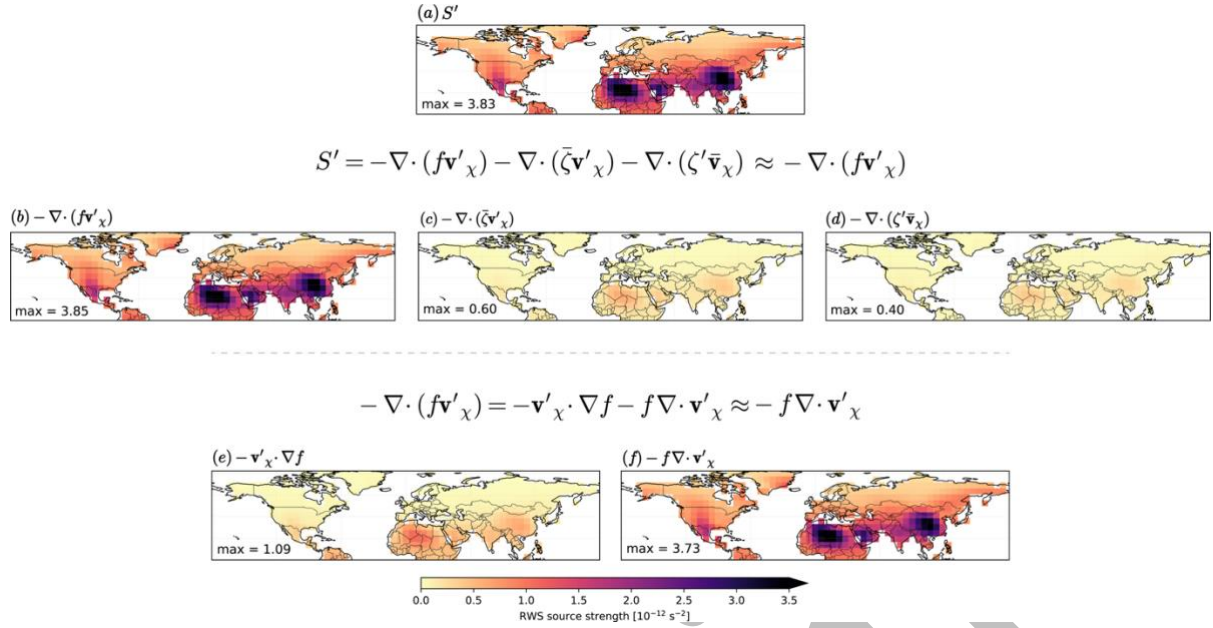


Fig. 5. Scale analysis of the strength of each term of Rossby wave source terms during the March–May season. Absolute values of each term at all available pressure levels are used for strength calculation, using the 3×3 grid cells surrounding the forcing location instead of a $15000\text{km} \times 7500\text{km}$ box. The maximum value of each term is shown in the bottom left corner of each panel. The equations related to displayed terms are shown between panels.

b. Reconstructed Response Strength

The seasonal response strength (Eq. 1) is reconstructed using the approximation for the absolute values of changed geopotential height derived in Eq. 8 (Fig. 6). Interestingly, this approximation qualitatively reproduces the spatial patterns of the original response strength shown in Fig. 2. In particular, the patterns in the midlatitudes, including the locations of the hotspots, are well reproduced.

However, minor differences are observed outside the midlatitudes. The magnitude of the reconstructed response strength is relatively weak at low latitudes, indicated by dark blue shading, and relatively large values in high latitude regions, where the reconstruction shows magnitudes like those found in low latitude regions in Fig. 2. These differences arise primarily from the assumption that approximated geopotential with streamfunction ($gz \approx f\psi$), which is most valid along the midlatitudes. Because the approximation between geopotential and streamfunction depends on the Coriolis parameter (f), the estimated magnitude of geopotential becomes very small near the equator and relatively large near the

poles. These biases in magnitude, however, do not substantially affect identification of the hotspots.

The seasonal fluctuation of magnitude at each location is also not reflected in the reconstructed R . While the magnitude of R is generally strongest in winter and weakest in summer in the original response strength, such seasonal fluctuation is not observed in the reconstructed R . This discrepancy is primarily associated with the assumption of a fixed wavelength (see Appendix for details), which is not valid across different seasons and locations. Allowing wavelength to vary spatially and seasonally could improve the representation of the seasonal amplitude. Nevertheless, applying variable wavelengths produce only minor changes in the overall spatial patterns of the reconstructed R (not shown).

Overall, Fig. 6 confirms that the reconstructed response strength successfully captures the major spatial features of the original response, particularly the locations of the hotspot regions identified in Fig. 2. Therefore, the approximation (Eq. 8) provides a useful framework for qualitatively investigating the physical processes that contribute to stronger atmospheric wave responses in the midlatitudes.

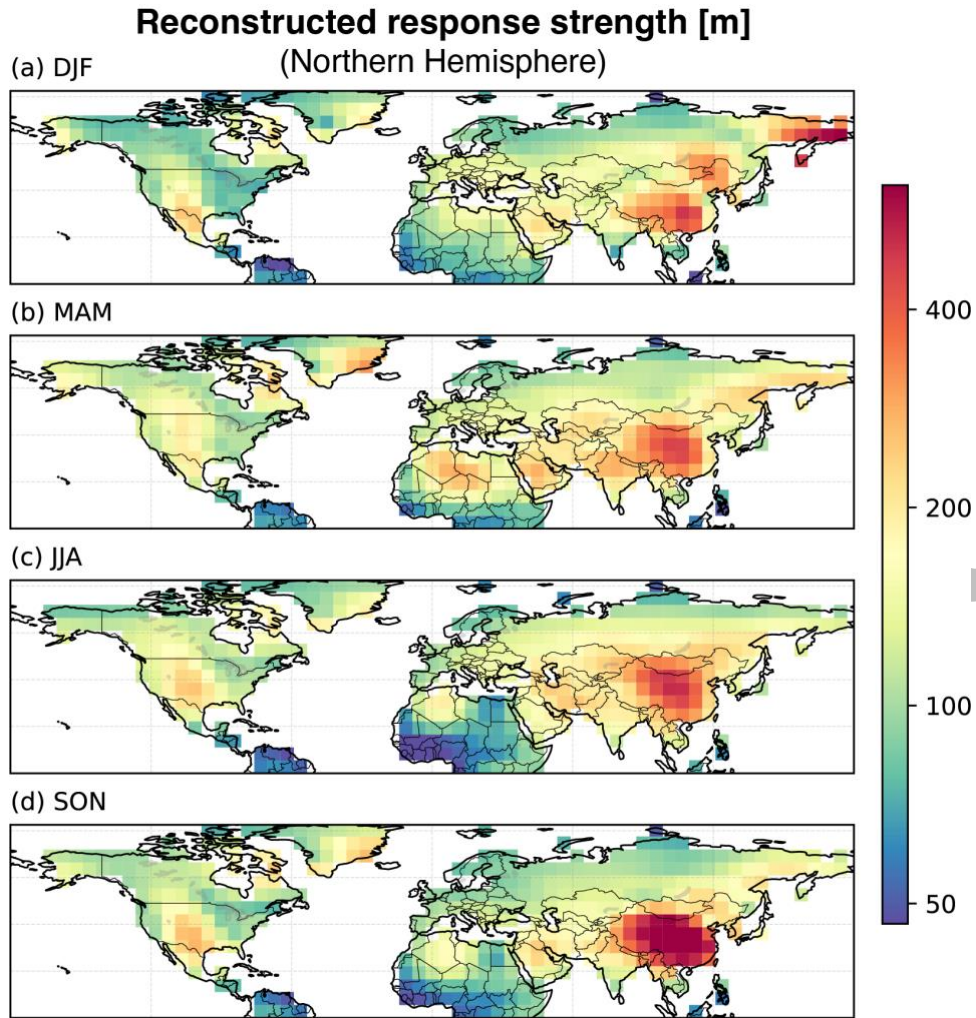


Fig. 6. Reconstructed Response Strength. Spatial patterns of reconstructed response strength are shown for each season, with shadings displayed on a log-scale colormap.

5. Conditions for Stronger Responses

Regarding that $\nabla \cdot \overrightarrow{v'_\chi}$ is the only term in Eq. 8 that depends directly on the response of each experiment, whereas the remaining terms vary primarily with latitude, we further investigate the conditions that favor stronger $\nabla \cdot \overrightarrow{v'_\chi}$ responses. In this section, we use the continuity and thermodynamic equations to identify the physical processes associated with enhanced $\nabla \cdot \overrightarrow{v'_\chi}$.

a. Estimating $\nabla \cdot \overrightarrow{v'_\chi}$.

From the continuity equation in pressure coordinates, the horizontal divergence at a given pressure level can be related to vertical velocity (ω), as

$$\nabla \cdot \overrightarrow{v'_\chi} + \frac{\partial \omega'}{\partial p} = 0.$$

Rearranging and vertically integrating the equation gives

$$\omega'(p) = - \int_0^p \nabla \cdot \overrightarrow{v'_\chi} dp. \quad (9)$$

Eq. 9 provides a link between the strength of vertical motion and the vertically integrated horizontal divergence. The equation shows that strong upward motion ($\omega' < 0$) implies substantial net divergence in the atmospheric column above ($\nabla \cdot \overrightarrow{v'_\chi} > 0$). As ω is directly available as an output of the LBM in pressure coordinates, we use the magnitude of ω' as a proxy for the strength of the induced divergent circulation.

The thermodynamic equation provides a framework for identifying the processes that contribute to stronger vertical motion with a prescribed diabatic heating anomaly:

$$\frac{\partial T}{\partial t} + V \cdot \nabla T + \omega S_p = \frac{Q}{c_p}$$

where T is temperature, $V = (u, v)$ is wind vector, $S_p = \frac{\partial T}{\partial p} - \frac{R}{c_p} \frac{T}{p}$ is the dry static stability, R here is the gas constant for dry air, c_p is the specific heat at constant pressure, and Q is the diabatic heating. Under the steady-state assumption, the linearized thermodynamic equation, after neglecting smaller terms, can be written as

$$V' \cdot \nabla \bar{T} + \bar{V} \cdot \nabla T' + \omega' \bar{S}_p = Q'/c_p.$$

Rearranging the equation gives

$$\omega' = \frac{1}{\bar{S}_p} \left(\frac{Q'}{c_p} - V' \cdot \nabla \bar{T} - \bar{V} \cdot \nabla T' \right). \quad (10)$$

For simplicity, the overbar on $\bar{S}_p$ is omitted hereafter, and S_p denotes the climatological dry static stability. Eq. 10 indicates that the induced vertical velocity is controlled by static stability, imposed diabatic heating, and horizontal temperature advection terms. Therefore, if we impose the same adiabatic heating everywhere, stronger vertical motion, and consequently stronger divergent circulation, is expected in regions with lower climatological stability or with enhanced temperature advection.

b. Conditions for Strong Vertical Motion ω'

1) STATIC STABILITY

The spatial pattern of seasonally strong perturbed vertical velocity from LBM results broadly captures the location of the identified response strength hotspots (Figs. 7, S8–10). Dry static stability is calculated from the seasonal climatological temperature, because the dry version of LBM is used. To represent the vertical motion associated with the local responses around each forcing location, ω' is averaged over the 3×3 grid cells surrounding the forcing location (Fig. 7a).

Strong vertical motions are induced across several midlatitude regions, including the Great Plains, Sahara region, and areas surrounding the Tibetan Plateau, as indicated by darker shading in Fig. 7. These regions correspond to the hotspots identified in Fig. 2. In addition, Fig. 7 shows that regions of strong vertical velocity are collocated with regions of relatively low dry static stability. Lower values of S_p , indicated by darker shadings, are often found near the regions where large ω' responses are observed, which is consistent with Eq. 10.

Some discrepancies exist near the Tibetan Plateau, where strong vertical motion is induced despite relatively high static stability. This indicates that static stability alone cannot explain the vertical velocity response and suggests that strong warm advection should be properly accounted for, this is examined in the following section. Together, Fig. 7 confirms that the regions with lower climatological dry static stability respond more strongly to an imposed heating anomaly, producing stronger vertical velocity and, in turn, stronger divergent circulation.

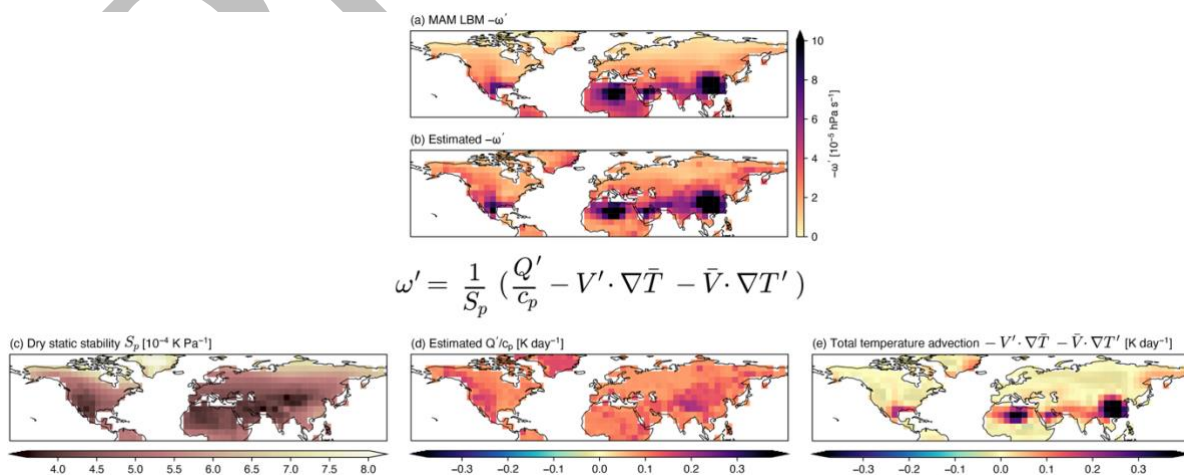


Fig. 7. Terms in the thermodynamic equation for the MAM season, averaged over 850–300hPa. (a) The LBM mean and (b) estimated vertical velocity, (c) dry static stability, (d) estimated diabatic heating, and (e) estimated total temperature advection averaged over 850–300 hPa. Levels closest to the surface are excluded from the calculation. The estimated

vertical velocity is calculated as the sum of diabatic heating and temperature advection divided by static stability.

2) TEMPERATURE ADVECTION

Further analysis reveals that the anomalous temperature advection terms account for much of the spatial structure of the induced vertical velocity response. The terms on the right-hand side of Eq. 10 are calculated for spring, including the estimated vertical velocity calculated using Eq. 10 (Fig. 7b), diabatic heating (Fig. 7d), and total horizontal temperature advection (Fig. 7e). The spatial pattern of ω' diagnosed from Eq. 10 (Fig. 7b) closely resembles the vertical velocity directly from the LBM in Fig. 7a. Again, the regions of strong vertical velocities are found near the response strength hotspots. This agreement allows the individual terms in Eq. 10 to be examined to determine which processes contribute to the spatial variations in vertical velocity responses.

By design, all experiments are forced with the same low-level diabatic heating anomaly in sigma coordinates (Fig. 1a), the native vertical coordinate of the LBM. While the imposed heating has the same vertical structure at a given sigma level across experiments, Fig. 7d shows heterogeneous spatial pattern of the heating averaged over 850–300 hPa. Elevated regions with lower basic state surface pressure are emphasized because the selected pressure level is closer to the prescribed low-level heating there. Overall, the imposed diabatic heating primarily sets the magnitude of the vertical-motion response rather than its spatial structure.

In contrast, the spatial patterns of the total temperature advection resemble the pattern of the diagnosed vertical velocity (Fig. 7e). Strong warm advection contributions are concentrated close to the regions of strong vertical velocity which also correspond to the identified hotspot regions. Outside these regions, the local temperature advection contribution is generally weaker and even shows cold advection with darker yellow shadings, which cannot explain induced vertical motions. Still, the magnitude of diabatic heating terms is typically larger than these weak cold temperature advection contributions, so that upward motion remains the dominant response to the imposed low-level heating. The same qualitative relationship between temperature advection and the spatial structure of the vertical motion response is found in the other seasons (Figs. S7–9).

The anomalous total temperature advection can be decomposed into two linear terms. One is the advection of the climatological temperature by the perturbed wind ($-V' \cdot \nabla \bar{T}$), and

the other is the advection of the perturbed temperature by the climatological wind ($-\bar{V} \cdot \nabla T'$). We first consider the structure of the climatological fields to simplify these terms. As the climatological temperature field mostly varies in the meridional direction, the climatological zonal temperature gradient ($\partial \bar{T} / \partial x$) is generally smaller than the meridional temperature gradient ($\partial \bar{T} / \partial y$). Thus, the first term can be approximated as

$$-V' \cdot \nabla \bar{T} \approx -v' \frac{\partial \bar{T}}{\partial y}. \quad (11)$$

The climatological meridional temperature gradient is generally stronger in winter and weaker in summer, reflecting the seasonal variability of large-scale horizontal temperature contrasts. Through thermal wind balance, stronger meridional temperature gradients are associated with stronger vertical shear of the zonal geostrophic wind and, in turn, stronger upper-level westerlies and climatological jet. Because the climatological zonal wind is typically much stronger than the meridional wind in the midlatitudes, the second anomalous advection term can similarly be approximated as

$$-\bar{V} \cdot \nabla T' \approx -\bar{u} \frac{\partial T'}{\partial x}. \quad (12)$$

Fig. 8 shows that advection of the climatological temperature by perturbed wind ($-V' \cdot \nabla \bar{T}$; Fig. 8a) accounts for most of the strong warm advection signal found in Fig. 7e. On the other hand, advection of perturbed temperature by climatological wind ($-\bar{V} \cdot \nabla T'$; Fig. 8b) opposes this response in several regions through cold advection. Eq. 11 suggests that strong warm advection is favored where strong perturbed meridional wind (v') coincides with strong climatological meridional temperature gradient ($\partial \bar{T} / \partial y$). The latter is dynamically related to the climatological jet through thermal wind balance, enhanced warm advection is expected within or near regions of strong background westerlies ($\bar{u}$). Consistent with this, Fig. S3b shows that regions of strong warm advection tend to overlap with regions of strong climatological zonal wind.

Stronger perturbed meridional winds (v' ; Fig. S11 right) are found near the regions of elevated terrain (Fig. 8c; Fig. S11). When the low-level heating is imposed over elevated terrain, the heating is located closer to a given pressure level than experiments over lower-elevated regions. As a result, stronger temperature perturbation can develop near elevated terrain at that pressure level (Fig. S11 left). The circulation associated with the induced thermal low (Fig. S10a) produces warm advection on the eastern side of the forcing and cold

advection on the western side at lower-level (Fig. S10b, c), increasing the zonal temperature gradient near the heating region (Fig. S10a). Thus, the steady state temperature response exhibits strong zonal temperature gradients at lower levels close to the imposed heating ($\partial T'/\partial x$; Fig. S11 middle). In turn, through thermal wind balance, these enhanced zonal temperature gradients are associated with stronger vertical wind shear of the perturbed meridional geostrophic wind. Consistent with this, enhanced perturbed upper-level meridional winds are found near the elevated forcing locations (v' ; Fig. S11 right).

These results suggest that topography can amplify the local temperature and circulation perturbations induced by prescribed low-level heating anomaly. Together, the background meridional temperature gradient associated with the climatological jets and the enhanced perturbation response near elevated terrain favor stronger anomalous warm advection and stronger vertical motion.

Similar factors also influence the second temperature advection term ($-\bar{V} \cdot \nabla T'$; Fig. 8b), but this term produces cold advection over much of the midlatitude lower troposphere (Fig. S12). From Eq. 12, its magnitude can be enhanced where strong climatological zonal winds ($\bar{u}$) coincide with strong perturbed zonal temperature gradient ($\partial T'/\partial x$). As discussed above, forcings imposed over elevated regions tend to produce stronger perturbed temperature gradients at a given pressure level. When combined with climatologically strong westerly winds, this typically yields anomalous cold advection in the lower troposphere in midlatitudes, consistent with the classical linear response to localized thermal forcing (Hoskins & Karoly, 1981). In the meantime, Fig. S12 shows that the regions of strong warm advection found in the Fig. 8b, such as in eastern Asia, arise primarily from the upper-level response, where the sign of the perturbed temperature gradient is reversed (Fig. S11).

To summarize, regions near the climatological jet and elevated terrain favor large magnitudes of both anomalous warm and cold temperature advection. However, within the response strength hotspots, warm advection associated with the perturbed wind acting on the climatological temperature ($-V' \cdot \nabla \bar{T}$) is dominant, resulting in stronger vertical velocity responses. The presence of topography serves as a springboard, strengthening the local temperature and upper-level meridional wind perturbation, whereas proximity to the climatological jet provides strong background meridional temperature gradient. The synergy between these two processes favors enhanced anomalous warm advection and stronger vertical motion through the thermodynamic balance.

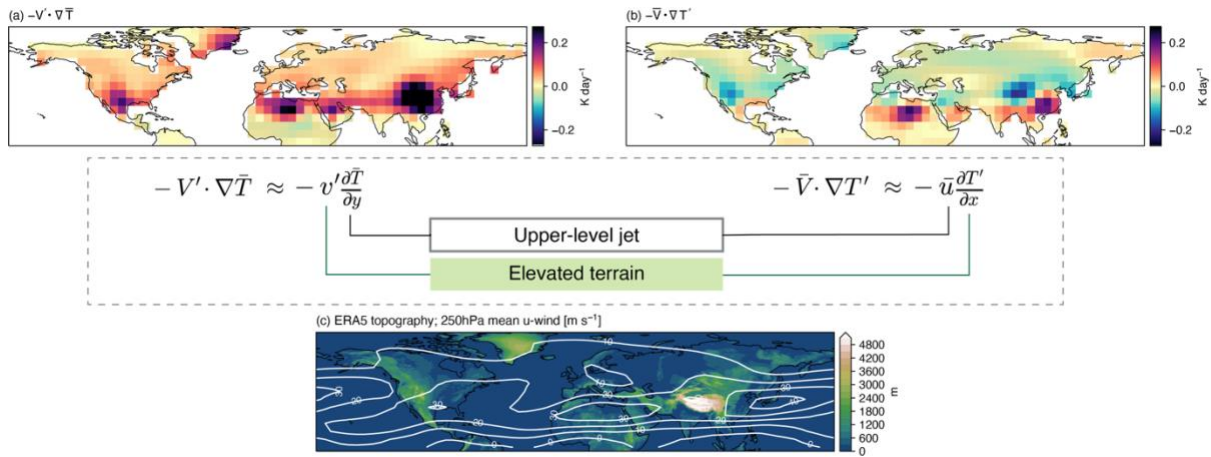


Fig. 8. Estimated temperature advection terms and associated winds, temperature gradient, and topography for the MAM season. Shown are (a) temperature advection of the climatological temperature by the perturbed wind, (b) temperature advection of the perturbed temperature by the climatological wind, and (c) topography from ERA5. Panels (a) and (b) show values averaged over 850-300hPa. Eqs. 11 and 12 are embedded to describe connections between shown terms.

6. Summary of Conditions Leading to Stronger Responses

Fig. 9 summarizes the processes that favor stronger atmospheric responses to imposed diabatic heating in the hotspot regions, highlighting contributions from both the mean background state (solid boxes) and induced perturbed state (dashed boxes). A stronger response strength reflects the magnitude of height changes in the atmosphere, which is linked to the strength of convergent/divergent motions throughout the atmospheric column. The induced vertical motion provides a useful measure of this column-integrated convergence and divergence.

Strong vertical motion responses are favored when the heating is imposed over the regions of relatively low climatological static stability (Box 1) and where anomalous warm advection is enhanced (Box 4). Strong warm advection is associated with two mean state characteristics: proximity to elevated terrain (Box 2) and climatological upper-level jet (Box 3). Forcings located over elevated terrain induces stronger local temperature and circulation perturbations, and regions near upper-level jet are characterized by strong climatological meridional temperature gradients. Together, these conditions favor stronger anomalous warm advection.

When low-level heating is imposed within these favorable environments, stronger low-level convergence develops near the forcing center (Box 5), accompanied by stronger vertical motion (Box 6). In turn, mass continuity requires a correspondingly stronger divergent response aloft. The upper-level divergence and associated positive geopotential height anomaly are displaced eastward or northeastward from the forcing center.

The eastward displacement and structure of the upper-level response are also related to the perturbed temperature field at the lower-level. Through thermal wind balance, vertical shear of perturbed meridional winds is enhanced near the forcing location. This perturbed meridional wind shapes the western flank of the upper-level positive height anomaly and establishes the featured eastward displacement.

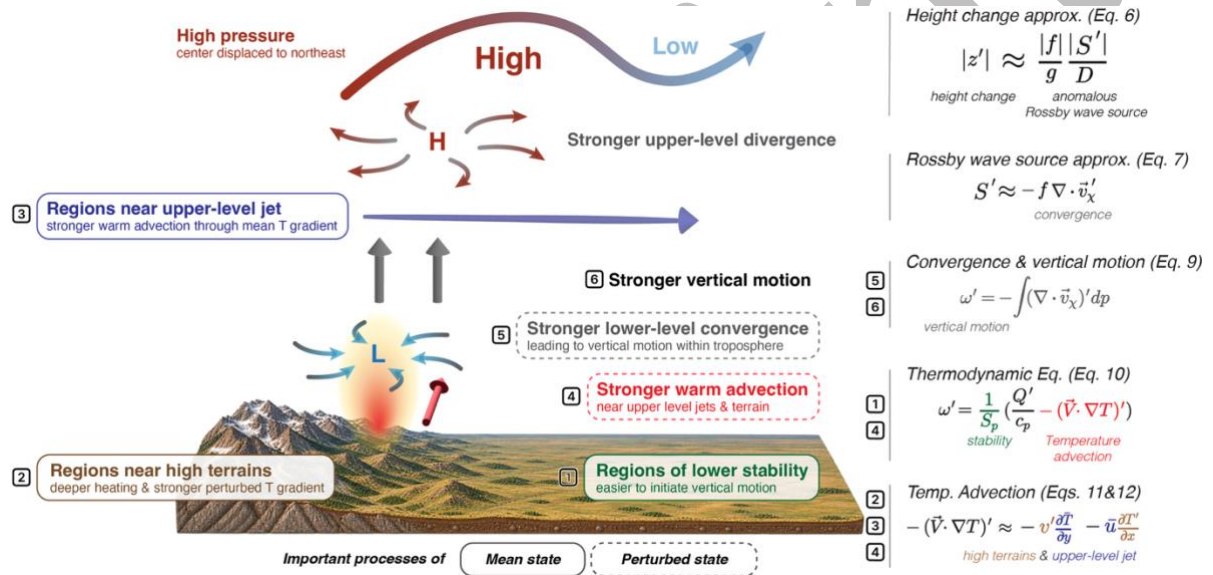


Fig. 9. Conceptual diagram of important processes of inducing stronger atmospheric responses due to imposed low-level heating anomaly. Important processes related to mean (perturbed) state are marked as boxes with solid (dashed) contour with number. Each number is also assigned to the related equations shown on the right.

7. Discussion and Conclusions

This study investigates how imposed low-level heating anomalies over land, which can arise due to persistent drought conditions, affect the large-scale atmospheric circulation. We analyze the physical mechanisms that make some regions produce stronger responses than others. We conducted approximately 2000 experiments using a Linear Baroclinic Model, taking advantage of its low computational cost due to its low complexity. The results show that regions near strong climatological upper-level zonal flow and topography, combined

with low static stability, exhibit stronger warm advection and vertical motion in response to low-level heating, leading to stronger perturbation in both lower and upper atmosphere.

While previous studies have conducted similar idealized heating experiments over selected regions, our study extends this framework to global land areas and examines the physical and dynamical processes that distinguish regions of particularly strong wave responses. Koster et al. (2016) and Wang et al. (2019) showed that prescribed diabatic heating can generate largely phase locked atmospheric circulation responses, suggesting a potential feedback pathway between dry land surface conditions and the large-scale circulation at certain regions. Here, we extend this approach globally and identified the general spatial structure and their seasonal dependence of the resulting wave responses. Further, we identify the factors that modulate the strength of the circulation response to drought-like diabatic heating.

A previous study showed that circulation anomalies, the development of upper-level high pressure system and enhanced lower-level divergence, can explain surface drying even in the absence of explicit information about atmospheric moisture (Kim et al., *submitted*). Our results suggest that if droughts persist within the hotspot regions, induced anomalous upper-level anticyclonic circulation, typically displaced northeast from the heating source, may favor drying in adjacent regions. This provides a potential dynamical pathway for the spatial propagation of drought. The 2012 US drought is one example of drought propagation. Persistent drought conditions affected the southern Great Plains during spring 2012, followed by drying over the Midwest located northeast in subsequent months (Hoerling et al., 2014; DeAngelis et al., 2020). Because the southern Great Plains is located within the hotspots identified here, the idealized results suggest that land-surface heating in this region may have contributed to circulation anomalies favorable for drying in the regions located northeast. More generally, hotspot regions may be particularly efficient in amplifying or propagating drought favorable conditions once persistent droughts are present over the regions. Furthermore, several hotspot regions, including near the Great Plains, Tibetan Plateau, and southern Africa are projected to experience increased atmospheric drying (Wankmüller et al., 2024), which can contribute to more widespread dry conditions in the future through the proposed pathway.

An important caveat is the missing physical link between dry land surface conditions and the imposed low-level heating anomaly. The conversion of soil moisture deficits into sensible

heating depends on land surface properties and background climate. Differences in soil properties affect the surface energy balances (Lehmann et al., 2018; Wankmüller et al., 2024; Dennis and Berbery, 2021), which can result in different warming responses across regions (Miralles et al., 2012; Hsu et al., 2024). In addition, the warming response is dependent on the season. For example, Teng et al. (2019) showed that different climatological soil state and planetary boundary layer structure across seasons can result in different heating responses to the same soil moisture anomalies. Thus, the underlying assumption in this study that an identical soil moisture anomaly would generate a uniform heating response across all land grid cells is unlikely to hold. The magnitude and vertical profile of the resulting temperature perturbation may vary considerably across regions and seasons.

The strength of the atmospheric response depends not only on the geographical location of the given heating anomaly, as examined here, but also on the efficiency of dry land surface conditions to generate that heating anomaly. Some regions or seasons may produce strong atmospheric responses even if they are not identified as dynamical hotspots in the present study, because they can generate stronger or deeper heating anomalies for a given soil moisture deficit. For example, regions such as the Mediterranean during summer can exhibit particularly strong land surface heating response to soil moisture anomalies in summertime (Miralles et al., 2012). A more complete assessment of drought-circulation feedback will therefore require consideration of both components – the sensitivity of atmospheric circulation to diabatic heating and the sensitivity of diabatic heating to land surface drying.

A second limitation is that the dry version of LBM used here does not resolve any processes involving atmospheric moisture, and nonlinear dynamical interactions are not considered by construction. Inclusion of these processes could modify the simulated responses examined in our study. Changes in moisture could affect horizontal and vertical temperature gradients and thereby influence the associated circulation. If more complex processes are considered, such as cloud and convective processes, it could further modify the response through latent heating, particularly where moist convection is active. Such effects are likely to be important over the tropical oceans. However, our goal is to understand how low-level heating associated with dry land surface conditions changes the atmospheric circulation, including such processes would not significantly change our conclusion. Moreover, the Koster et al. (2016) study already compared responses from a simple

linearized model with those from a more complex atmospheric general circulation model and found broadly consistent large-scale circulation responses.

This study investigated a thermodynamic and dynamical pathway linking land surface-related low-level heating anomalies to large-scale atmospheric circulation responses, exploring a potential pathway of the remote effect of land-atmosphere interactions. The results suggest that drought amplification and spatial propagation may be particularly favored when heating occurs over persistent dry land surface in regions where the background atmospheric state efficiently converts that heating into a large-scale circulation response. However, the pathway linking soil moisture anomalies to sufficient low-level heating anomalies remains unsolved and requires further investigation. Future studies using a hierarchy of models could explicitly couple these two components and evaluate to what extent anomalies over the hotspot regions contribute to the amplification and spatial propagation of historical drought events.

Acknowledgments.

Paste your acknowledgments here.

Data Availability Statement.

The LBM source code used in this study is available from their website (<https://ccsr.aori.u-tokyo.ac.jp/~lbm/sub/lbm.html>).

APPENDIX

Approximation of the Streamfunction

The perturbed streamfunction can be approximated locally as a wave-like solution,

$$\psi' \approx \psi_0 e^{i(kx+ly)},$$

where $k = 2\pi/L_x$ and $l = 2\pi/L_y$ are the zonal and meridional wavenumbers, respectively, and L_x and L_y are the corresponding wavelengths. We further define $K^2 = k^2 + l^2$. Under this approximation, the rotational wind perturbation is

$$v'_\psi = \left(-\frac{\partial \psi'}{\partial y}, \frac{\partial \psi'}{\partial x} \right) = (-il, ik)\psi',$$

and the relative vorticity perturbation is

$$\zeta' = \nabla^2 \psi' = (-k^2 - l^2)\psi' = -K^2 \psi'.$$

Therefore, Eq. 4 can be written as

$$(\overline{u_\psi}, \overline{v_\psi}) \cdot (ik, il)(-K^2 \psi') + (-il, ik)\psi' \cdot \nabla(\bar{\zeta} + f) = S'.$$

Arranging terms for ψ' gives

$$i[-K^2(\overline{u_\psi}k + \overline{v_\psi}l) - l\partial_x \bar{\zeta}_a + k\partial_y \bar{\zeta}_a]\psi' = S', \quad (A1)$$

where $\bar{\zeta}_a$ is the climatological absolute vorticity. Eq. A1 can be expressed as

$$i\delta\psi' = S',$$

where

$$\delta = \delta_{jet} + \delta_{vort},$$

with

$$\delta_{jet} \equiv -K^2(\overline{u_\psi}k + \overline{v_\psi}l)$$

$$\delta_{vort} \equiv -l\partial_x \bar{\zeta}_a + k\partial_y \bar{\zeta}_a.$$

We first assume constant zonal and meridional wavelengths, $L_x = L_y = 6000km$. Under this assumption, k and l are fixed, such that δ depends only on the background climatological state rather than the simulated response in each experiment.

To assess the sensitivity of the approximation to this assumption, Fig. S13 shows reconstructed response strength maps obtained using seasonally varying stationary zonal wavenumbers. Specifically, we prescribed $n_{DJF} = 5$, $n_{MAM} = 6$, $n_{JJA} = 7$, and $n_{SON} = 6$, and calculated the local zonal wavenumber as $k = n/a \cos \phi$, where a is Earth's radius and ϕ is latitude. Allowing the zonal wavenumber to vary produces spatial patterns of reconstructed response strength that are qualitatively similar to those obtained with the fixed wavelength assumption in Fig. 6. Thus, the choice of a fixed wavelength does not substantially affect the qualitative interpretation of the responses.

This approximation relates the streamfunction (ψ') directly to the anomalous RWS (S').
To prevent ψ' to blow up, we introduce a linear damping on the right-hand side of the Eq.
A1,

$$i\delta\psi' = S' - r\zeta',$$

where r is the linear damping coefficient. Using $\zeta' = -iK^2\psi'$ and rearranging yields

$$(i\delta - rK^2)\psi' \approx S',$$

and therefore

$$\psi' \approx \frac{S'}{i\delta - rK^2} = \frac{S'}{D}. \quad (A2)$$

As our analysis focuses on the amplitude of the streamfunction response, we consider the
magnitude of Eq. A2,

$$|\psi'| \approx \frac{|S'|}{|D|}. \quad (A3)$$

REFERENCES

- 793
- 794 Betts, A. K., Ball, J. H., Beljaars, A. C. M., Miller, M. J., & Viterbo, P. A. (1996). The land
795 surface-atmosphere interaction: A review based on observational and global modeling
796 perspectives. *J. Geophys. Res. Atmos.*, 101(D3), 7209–7225.
797 <https://doi.org/10.1029/95JD02135>
- 798 Branstator, G. (2002). Circumglobal Teleconnections, the Jet Stream Waveguide, and the
799 North Atlantic Oscillation. *J. Climate*, 15(14), 1893–1910. [https://doi.org/10.1175/1520-
800 0442\(2002\)015%253C1893:CTTJSW%253E2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015%253C1893:CTTJSW%253E2.0.CO;2)
- 801 Cook, B. I., Bonan, G. B., & Levis, S. (2006). Soil Moisture Feedbacks to Precipitation in
802 Southern Africa. *J. Climate*, 19(17), 4198–4206. <https://doi.org/10.1175/JCLI3856.1>
- 803 Coumou, D., Petoukhov, V., Rahmstorf, S., Petri, S., & Schellnhuber, H. J. (2014). Quasi-
804 resonant circulation regimes and hemispheric synchronization of extreme weather in
805 boreal summer. *Proc. Natl. Acad. Sci.*, 111(34), 12331–12336.
806 <https://doi.org/10.1073/pnas.1412797111>
- 807 DeAngelis, A. M., Wang, H., Koster, R. D., Schubert, S. D., Chang, Y., & Marshak, J.
808 (2020). Prediction Skill of the 2012 U.S. Great Plains Flash Drought in Subseasonal
809 Experiment (SubX) Models. *J. Climate*, 33(14), 6229–6253.
810 <https://doi.org/10.1175/JCLI-D-19-0863.1>
- 811 Dennis, E. J., & Berbery, E. H. (2021). The Role of Soil Texture in Local Land Surface–
812 Atmosphere Coupling and Regional Climate. *J. Hydrometeor.*, 22(2), 313–330.
813 <https://doi.org/10.1175/JHM-D-20-0047.1>
- 814 Fischer, E. M., Seneviratne, S. I., Lüthi, D., & Schär, C. (2007). Contribution of land-
815 atmosphere coupling to recent European summer heat waves. *Geophys. Res. Lett.*, 34(6),
816 2006GL029068. <https://doi.org/10.1029/2006GL029068>
- 817 Guillod, B. P., Orlowsky, B., Miralles, D. G., Teuling, A. J., & Seneviratne, S. I. (2015).
818 Reconciling spatial and temporal soil moisture effects on afternoon rainfall. *Nat.*
819 *Commun.*, 6(1), 6443. <https://doi.org/10.1038/ncomms7443>
- 820 Hoerling, M., Eischeid, J., Kumar, A., Leung, R., Mariotti, A., Mo, K., Schubert, S., &
821 Seager, R. (2014). Causes and Predictability of the 2012 Great Plains Drought. *Bull.*
822 *Amer. Meteor. Soc.*, 95(2), 269–282. <https://doi.org/10.1175/BAMS-D-13-00055.1>

- 823 Hoskins, B., & Woollings, T. (2015). Persistent Extratropical Regimes and Climate
824 Extremes. *Curr. Clim. Change Rep.*, 1(3), 115–124. [https://doi.org/10.1007/s40641-015-](https://doi.org/10.1007/s40641-015-0020-8)
825 [0020-8](https://doi.org/10.1007/s40641-015-0020-8)
- 826 Hsu, H., Dirmeyer, P. A., & Seo, E. (2024). Exploring the Mechanisms of the Soil Moisture-
827 Air Temperature Hypersensitive Coupling Regime. *Water Resour. Res.*, 60(7),
828 e2023WR036490. <https://doi.org/10.1029/2023WR036490>
- 829 Jenney, A. M., Randall, D. A., & Barnes, E. A. (2021). Drivers of uncertainty in future
830 projections of Madden–Julian Oscillation teleconnections. *Weather Clim. Dyn.*, 2(3),
831 653–673. <https://doi.org/10.5194/wcd-2-653-2021>
- 832 Jiang, J., Liu, Y., Meng, J., Wu, G., He, B., Ma, T., Bao, W., & Fan, J. (2024). Dry soil
833 moisture on the Tibetan plateau drives synchronous extreme heatwaves in Europe and
834 East Asia. *npj Clim. Atmos. Sci.*, 7(1), 281. <https://doi.org/10.1038/s41612-024-00831-x>
- 835 Kornhuber, K., Coumou, D., Vogel, E., Lesk, C., Donges, J. F., Lehmann, J., & Horton, R.
836 M. (2020). Amplified Rossby waves enhance risk of concurrent heatwaves in major
837 breadbasket regions. *Nat. Clim. Change*, 10(1), 48–53. [https://doi.org/10.1038/s41558-](https://doi.org/10.1038/s41558-019-0637-z)
838 [019-0637-z](https://doi.org/10.1038/s41558-019-0637-z)
- 839 Kornhuber, K., Petoukhov, V., Karoly, D., Petri, S., Rahmstorf, S., & Coumou, D. (2017).
840 Summertime Planetary Wave Resonance in the Northern and Southern Hemispheres. *J.*
841 *Climate*, 30(16), 6133–6150. <https://doi.org/10.1175/JCLI-D-16-0703.1>
- 842 Koster, R. D., Chang, Y., & Schubert, S. D. (2014). A Mechanism for Land–Atmosphere
843 Feedback Involving Planetary Wave Structures. *J. Climate*, 27(24), 9290–9301.
844 <https://doi.org/10.1175/JCLI-D-14-00315.1>
- 845 Koster, R. D., Chang, Y., Wang, H., & Schubert, S. D. (2016). Impacts of Local Soil
846 Moisture Anomalies on the Atmospheric Circulation and on Remote Surface
847 Meteorological Fields during Boreal Summer: A Comprehensive Analysis over North
848 America. *J. Climate*, 29(20), 7345–7364. <https://doi.org/10.1175/JCLI-D-16-0192.1>
- 849 Koster, R. D., Dirmeyer, P. A., Guo, Z., Bonan, G., Chan, E., Cox, P., Gordon, C. T., Kanae,
850 S., Kowalczyk, E., Lawrence, D., Liu, P., Lu, C.-H., Malyshev, S., McAvaney, B.,
851 Mitchell, K., Mocko, D., Oki, T., Oleson, K., Pitman, A., ... Yamada, T. (2004).

- 852 Regions of Strong Coupling Between Soil Moisture and Precipitation. *Science*,
853 305(5687), 1138–1140. <https://doi.org/10.1126/science.1100217>
- 854 Lehmann, P., Merlin, O., Gentine, P., & Or, D. (2018). Soil Texture Effects on Surface
855 Resistance to Bare-Soil Evaporation. *Geophys. Res. Lett.*, 45(19).
856 <https://doi.org/10.1029/2018GL078803>
- 857 Lu, R., & Lin, Z. (2009). Role of Subtropical Precipitation Anomalies in Maintaining the
858 Summertime Meridional Teleconnection over the Western North Pacific and East Asia.
859 *J. Climate*, 22(8), 2058–2072. <https://doi.org/10.1175/2008JCLI2444.1>
- 860 Martius, O., Wehrli, K., & Rohrer, M. (2021). Local and Remote Atmospheric Responses to
861 Soil Moisture Anomalies in Australia. *J. Climate*, 34(22), 9115–9131.
862 <https://doi.org/10.1175/JCLI-D-21-0130.1>
- 863 Miralles, D. G., Gentine, P., Seneviratne, S. I., & Teuling, A. J. (2019). Land–atmospheric
864 feedbacks during droughts and heatwaves: State of the science and current challenges.
865 *Ann. N. Y. Acad. Sci.*, 1436(1), 19–35. <https://doi.org/10.1111/nyas.13912>
- 866 Miralles, D. G., Van Den Berg, M. J., Teuling, A. J., & De Jeu, R. A. M. (2012). Soil
867 moisture-temperature coupling: A multiscale observational analysis. *Geophys. Res. Lett.*,
868 39(21), 2012GL053703. <https://doi.org/10.1029/2012GL053703>
- 869 Petoukhov, V., Rahmstorf, S., Petri, S., & Schellnhuber, H. J. (2013). Quasiresonant
870 amplification of planetary waves and recent Northern Hemisphere weather extremes.
871 *Proceedings of the National Academy of Sciences*, 110(14), 5336–5341.
872 <https://doi.org/10.1073/pnas.1222000110>
- 873 Santanello, J. A., Dirmeyer, P. A., Ferguson, C. R., Findell, K. L., Tawfik, A. B., Berg, A.,
874 Ek, M., Gentine, P., Guillod, B. P., Van Heerwaarden, C., Roundy, J., & Wulfmeyer, V.
875 (2018). Land–Atmosphere Interactions: The LoCo Perspective. *Bull. Amer. Meteor.*
876 *Soc.*, 99(6), 1253–1272. <https://doi.org/10.1175/BAMS-D-17-0001.1>
- 877 Sardeshmukh, P. D., Hoskings, B. J. (1988). The generation of global rotational flow by
878 steady idealized tropical divergence. *J. Atmos. Sci.*, 45(7), 1228–1251.
- 879 Schubert, S. D., Wang, H., Koster, R. D., Suarez, M. J., & Groisman, P. Ya. (2014). Northern
880 Eurasian Heat Waves and Droughts. *J. Climate*, 27(9), 3169–3207.
881 <https://doi.org/10.1175/JCLI-D-13-00360.1>

- 882 Schubert, S., Wang, H., & Suarez, M. (2011). Warm Season Subseasonal Variability and
883 Climate Extremes in the Northern Hemisphere: The Role of Stationary Rossby Waves.
884 *J. Climate*, 24(18), 4773–4792. <https://doi.org/10.1175/JCLI-D-10-05035.1>
- 885 Schumacher, D. L., Keune, J., Dirmeyer, P., & Miralles, D. G. (2022). Drought self-
886 propagation in drylands due to land–atmosphere feedbacks. *Nat. Geosci.*, 15(4), 262–
887 268. <https://doi.org/10.1038/s41561-022-00912-7>
- 888 Sena, A. C. T., & Williams, I. N. (2026). Understanding Drought-Induced Soil Moisture
889 Feedback on Atmospheric Circulation: Observations and Simulation Experiments of the
890 2020 Western U.S. Drought. *J. Geophys. Res. Atmos.*, 131(12), e2025JD044617.
891 <https://doi.org/10.1029/2025JD044617>
- 892 Seneviratne, S. I., Corti, T., Davin, E. L., Hirschi, M., Jaeger, E. B., Lehner, I., Orlowsky, B.,
893 & Teuling, A. J. (2010). Investigating soil moisture–climate interactions in a changing
894 climate: A review. *Earth-Sci. Rev.*, 99(3–4), 125–161.
895 <https://doi.org/10.1016/j.earscirev.2010.02.004>
- 896 Taylor, C. M., De Jeu, R. A. M., Guichard, F., Harris, P. P., & Dorigo, W. A. (2012).
897 Afternoon rain more likely over drier soils. *Nature*, 489(7416), 423–426.
898 <https://doi.org/10.1038/nature11377>
- 899 Teng, H., & Branstator, G. (2019). Amplification of Waveguide Teleconnections in the
900 Boreal Summer. *Curr. Clim. Change Rep.*, 5(4), 421–432.
901 <https://doi.org/10.1007/s40641-019-00150-x>
- 902 Teng, H., Branstator, G., Tawfik, A. B., & Callaghan, P. (2019). Circumglobal Response to
903 Prescribed Soil Moisture over North America. *J. Climate*, 32(14), 4525–4545.
904 <https://doi.org/10.1175/JCLI-D-18-0823.1>
- 905 Teng, H., Branstator, G., Wang, H., Meehl, G. A., & Washington, W. M. (2013). Probability
906 of US heat waves affected by a subseasonal planetary wave pattern. *Nat. Geosci.*, 6(12),
907 1056–1061. <https://doi.org/10.1038/ngeo1988>
- 908 Wang, H., Schubert, S. D., Koster, R. D., & Chang, Y. (2019). Phase Locking of the Boreal
909 Summer Atmospheric Response to Dry Land Surface Anomalies in the Northern
910 Hemisphere. *J. Climate*, 32(4), 1081–1099. <https://doi.org/10.1175/JCLI-D-18-0240.1>

- 911 Wankmüller, F. J. P., Delval, L., Lehmann, P., Baur, M. J., Cecere, A., Wolf, S., Or, D.,
912 Javaux, M., & Carminati, A. (2024). Global influence of soil texture on ecosystem water
913 limitation. *Nature*, 635(8039), 631–638. <https://doi.org/10.1038/s41586-024-08089-2>
- 914 Watanabe, M., & Jin, F.-F. (2004). Dynamical Prototype of the Arctic Oscillation as
915 Revealed by a Neutral Singular Vector. *J. Climate*, 17(11), 2119–2138.
916 [https://doi.org/10.1175/1520-0442\(2004\)017%253C2119:DPOTAO%253E2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017%253C2119:DPOTAO%253E2.0.CO;2)
- 917 Watanabe, M., & Kimoto, M. (2000). Atmosphere-ocean thermal coupling in the North
918 Atlantic: A positive feedback. *Q. J. R. Meteorol. Soc.*, 126(570), 3343–3369.
919 <https://doi.org/10.1002/qj.49712657017>
- 920