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To cite this article: Ning Wang *et al* 2026 *Environ. Res. Lett.* **21** 184020

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Atmospheric thickness links circum-hemisphere teleconnection to extreme cold events in boreal winter

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E-mail: ljp@ouc.edu.cn**Keywords:** extreme cold events, local atmospheric processes, atmospheric thickness, perturbation hypsometric law, circum-hemisphere teleconnectionSupplementary material for this article is available [online](#)**Abstract**

Extreme cold events (ECEs) impose societal risks, yet the local process linking large-scale teleconnections to regional ECEs is unclear. Here we systematically evaluate local atmospheric factors to identify the factor most strongly associated with the linkage between the boreal winter circum-hemisphere teleconnection and regional ECEs. Using a five-part statistical framework, we compare 1000–500 hPa thickness (ΔZ) with 13 other factors. ΔZ exhibits the strongest and most robust statistical association with ECEs among the factors examined: removing ΔZ produces the largest change in the ECE-related cold-tail distribution, $P(\text{ECEs} | \text{extreme } \Delta Z)$ is the highest among all factors, and the ΔZ –ECE relationship remains robust across multiple statistical analyses. The perturbation hypsometric law and complementary dynamical diagnostics provide a physical interpretation of this statistical relationship. Together, these results suggest that atmospheric thickness provides a physically interpretable local bridge from large-scale teleconnection signals to ECEs.

1. Introduction

Extreme cold events (ECEs) (see table S1 for a full list of abbreviations) remain high-impact winter hazards for health, transport, energy demand, and ecosystems (Quiroz 1984, Huynen *et al* 2001, Medina-Ramón and Schwartz 2007, Chen and Sun 2018, He *et al* 2023). Recent winters have still produced record-breaking cold episodes over Northern Hemisphere lands (Zhang *et al* 2022, Ye *et al* 2025). Moreover, although weak ECEs have generally declined under global warming, strong ECEs show no comparable decrease and may even increase in some regions (He *et al* 2023). Understanding the variability and causes of these persistent strong cold extremes is therefore crucial for improving winter climate-risk assessment and prediction.

Attribution of such extremes is challenging due to their multifaceted and complex nature (Trenberth *et al* 2015, Stott *et al* 2016). Many studies have linked regional winter ECEs to remote oceanic and

atmospheric drivers, which can be broadly grouped into four categories. First, Arctic sea ice loss and snow cover changes influence mid-latitude cold extremes through stratospheric pathways and polar vortex modulation (e.g. Mori *et al* 2014, Kug *et al* 2015, Matsumura and Kosaka 2019, Ye *et al* 2025). Second, tropical Pacific SST variability, including ENSO and the Victoria Mode, affects cold events over Eurasia and North America via planetary wave teleconnections (e.g. Rudeva and Simmonds 2021, Zheng *et al* 2022). Third, North Atlantic processes, such as the North Atlantic Oscillation, Atlantic Meridional Overturning Circulation, and the North Atlantic Tripole (NAT), modulate cold spells over Europe and East Asia (e.g. Yao *et al* 2017, Cohen *et al* 2021, Li *et al* 2022, Meccia *et al* 2024, Wang *et al* 2025). Finally, cross-basin synergistic effects, such as the combined influence of the North Pacific and North Tropical Atlantic SST, can further amplify cold event risks (e.g. Wang *et al* 2022, Zhang *et al* 2022, Ji *et al* 2025). These drivers can organize planetary-wave trains and



OPEN ACCESS

RECEIVED

23 July 2026

REVISED

30 August 2026

ACCEPTED FOR PUBLICATION

4 September 2026

PUBLISHED

18 September 2026

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circulation anomalies that favor cold extremes, but they do not identify the local atmospheric factor through which teleconnection signals most directly affect regional surface-temperature extremes. Several studies have noted local processes such as vertical motion (e.g. Song and Wu 2017, Gong *et al* 2024) and large-scale circulation systems (e.g. Bueh *et al* 2022, Dai *et al* 2025, Yu *et al* 2025), but without a systematic quantitative comparison of multiple candidate factors. For example, Wang *et al* (2023) considered surface albedo, water vapor, clouds, aerosols, surface fluxes, surface dynamics, and atmospheric dynamics terms, but did not isolate a concrete, measurable local driver with clear physical interpretation. Thus, a concise, measurable, and physically coherent local bridge between hemispheric teleconnections and regional ECEs remains lacking.

The circum-hemisphere teleconnection of extreme cold events (CHTe) describes five centers of action of winter cold night (TN10p) variability across Northern Hemisphere lands (Wang *et al* 2024). A related circum-hemisphere teleconnection (CHT), excited by the NAT, provides the large-scale circulation background for these cold extremes (Wang *et al* 2025). What remains unresolved is the local atmospheric factor through which the CHT influences regional ECEs.

Motivated by recent evidence that atmospheric thickness plays a key hydrostatic role in heat extremes over the Yangtze River basin (Sun *et al* 2025), we include atmospheric thickness between 1000 and 500 hPa (ΔZ) as a candidate local factor for ECEs. It is defined as $\Delta Z = Z_{500} - Z_{1000}$, with its anomaly given by $\Delta Z' = Z_{500}' - Z_{1000}'$, where Z_{500} and Z_{1000} denote the geopotential heights at 500 and 1000 hPa, respectively. In physical terms, ΔZ (m) represents the vertical separation between these two pressure surfaces. In this study, we propose the perturbation hypsometric law and use a multidimensional statistical framework to evaluate 14 local atmospheric factors. The statistical analyses identify ΔZ as the local atmospheric variable exhibiting the strongest and most robust association with regional ECEs among the factors examined, while the theoretical and dynamical analyses provide a physical interpretation of this statistical relationship.

2. Data and methods

2.1. Data

Detailed information on the datasets used in this study is provided in table S2. We use a monthly indicator for ECEs during 1901–2018 on a $1.25^\circ \times 1.875^\circ$ grid: cold nights (TN10p, percentage of time when daily minimum temperature <10th percentile), which is derived from the daily, *in situ* observations and supported by the World Meteorological Organization (Dunn *et al* 2020). All atmospheric

variables are obtained from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) Reanalysis 1 dataset (Kalnay *et al* 1996), including geopotential height, wind components, vertical velocity (ω), air temperature, sea level pressure, surface net latent heat flux, net sensible heat flux, net longwave radiation flux, and net shortwave radiation flux. Pressure-level variables are provided on a $2.5^\circ \times 2.5^\circ$ grid and surface flux variables on a $1.875^\circ \times 1.889^\circ$ grid, with both monthly and daily resolutions from January 1948 to May 2026. Considering the availability of both the TN10p and atmospheric datasets, the same study period (1951–2017) is selected. Anomalies are calculated as departures from the monthly climatology.

2.2. Indices

Winter refers to December–February (DJF). Local variables are averaged over the five CHTe regions following the CHTe index (CHTeI) definition of Wang *et al* (2024). Specifically, the CHTeI is calculated as a linear combination of normalized TN10p anomalies over the five centers of action:

$$\begin{aligned} \text{CHTeI} &= -\frac{1}{6} ([\text{TN10pa}_{R_1}] + [\text{TN10pa}_{R_3}] + [\text{TN10pa}_{R_5}]) \\ &\quad + \frac{1}{4} ([\text{TN10pa}_{R_2}] + [\text{TN10pa}_{R_4}]), \end{aligned} \quad (1)$$

where the square brackets denote the standardized area-mean TN10pa anomaly over the corresponding region. R_1 denotes Southeastern North America (78° – 93° W, 27° – 37° N and 78° – 108° W, 17° – 27° N), R_2 the Baffin Bay Coast (38° – 91° W, 46° – 78° N), R_3 Northern Europe (27° W– 53° E, 54° – 79° N), R_4 the Middle East–North Africa (22° – 68° E, 22° – 42° N and 16° W– 22° E, 22° – 32° N), and R_5 Eastern Siberia (82° – 140° E, 41° – 72° N).

ECEs and extreme local atmospheric factor events are defined as values below the 10th percentile of their respective distributions. The 14 tested factors are atmospheric thickness between 1000 and 500 hPa (ΔZ), geopotential height at 500 hPa (Z_{500}), geopotential height at 200 hPa (Z_{200}), sea level pressure (SLP), zonal wind at 200 hPa (U_{200}), meridional wind at 200 hPa (V_{200}), zonal wind at 850 hPa (U_{850}), meridional wind at 850 hPa (V_{850}), vertical motion at 500 hPa (ω_{500}), horizontal wind divergence at 200 hPa (DIV_{200}), net shortwave radiation flux (SR), net longwave radiation flux (LR), latent heat net flux (LHF), and sensible heat net flux (SHF). It should be noted that the 14 variables examined here do not represent equivalent physical quantities. Atmospheric thickness is directly related to layer-mean air temperature under hydrostatic balance, whereas the other local atmospheric factors primarily characterize dynamical

and thermodynamical processes that affect temperature tendencies through pathways such as temperature advection, vertical motion, and diabatic heating. Accordingly, the statistical analyses are intended to identify the local diagnostic quantity exhibiting the most stable and direct statistical relationship with ECEs, rather than to rank the causal effects of these variables.

2.3. Statistical methods

Surface air temperature anomaly (SATa) is first linearly regressed onto a given local atmospheric factor $x(t)$:

$$\widetilde{\text{SATa}}(t) = Ax(t) + B, \quad (2)$$

where $\widetilde{\text{SATa}}(t)$ represents the SATa component linearly associated with the local factor $x(t)$, and A and B are the regression coefficient and intercept, respectively. This linearly associated regression component is then removed from the original SATa(t) to obtain the residual temperature series SATa*(t):

$$\text{SATa}(t)^* = \text{SATa}(t) - \widetilde{\text{SATa}}(t), \quad (3)$$

Thus, SATa*(t) represents SATa after statistically removing the linear component associated with the local factor $x(t)$. This residualization should not be interpreted as completely removing the physical or causal effect of that factor.

To identify which local atmospheric factor exhibits the strongest statistical association with ECEs, we compare the probability density functions (PDFs) of SATa with those of the SATa*, as illustrated in figure S1. Both SATa and SATa* are normalized to have zero mean and unit variance before PDF estimation.

The cold tail is defined as the region below the 10th percentile of the SATa distribution (denoted as μ in figure S1), consistent with the definition of ECEs used throughout this study. We then compute the $\Delta\text{PDF} = \text{PDF}_{\text{SATa}^*} - \text{PDF}_{\text{SATa}}$ and interpret three patterns: (1) $\Delta\text{PDF} > 0$ in the cold tail (red curve in figure S1): Removing factor f increases the probability density in the cold tail relative to the SATa, indicating a negative statistical association between factor f and the ECE-related cold-tail distribution in this diagnostic. (2) $\Delta\text{PDF} < 0$ in the cold tail (blue curve): Removing factor f decreases the cold-tail probability density, indicating a positive statistical association between factor and the ECE-related cold-tail distribution. (3) $\Delta\text{PDF} = 0$ throughout the cold tail (black dashed curve): Removing factor f has negligible impact on the cold-tail distribution, indicating a weak statistical association in this diagnostic.

The remaining statistical diagnostics include tail Kullback–Leibler (K–L) and Jensen–Shannon (J–S) divergences for cold-tail distributional changes (text S1), contingency-table analysis for co-occurrence and

association strength (text S2), paired contingency analysis for changes in ECE occurrence after residualization (text S3), and the extreme consistency rate for lead–lag matching between extreme local factors and ECEs (text S4).

Other statistical methods, such as detrended analysis, and partial correlation are also used.

2.4. Dynamical METHODS

To provide a physical basis for using atmospheric thickness as a local pathway, we propose the perturbation hypsometric law (figure 1). This law is not an independent physical law but follows directly from the classical hypsometric relationship applied to both basic and perturbed hydrostatic states. Previous studies have mainly focused on the relationship between atmospheric thickness and layer-mean air temperature or their perturbations separately. Here, we further establish a unified framework that explicitly links the mean and perturbation states.

For an air column bounded by two fixed isobaric surfaces p_1 and p_2 , the hypsometric equation (Holton and Hakim 2013) gives

$$\Delta Z = Z_2 - Z_1 = \frac{R\langle T \rangle}{g_0} \ln \frac{p_1}{p_2}, \quad (4)$$

where ΔZ is the geopotential thickness, $\langle T \rangle$ is the layer-mean temperature, R is the gas constant for dry air, and g_0 is gravitational acceleration. Thus, for fixed p_1 and p_2 , ΔZ and $\langle T \rangle$ are constrained by a constant hydrostatic proportionality.

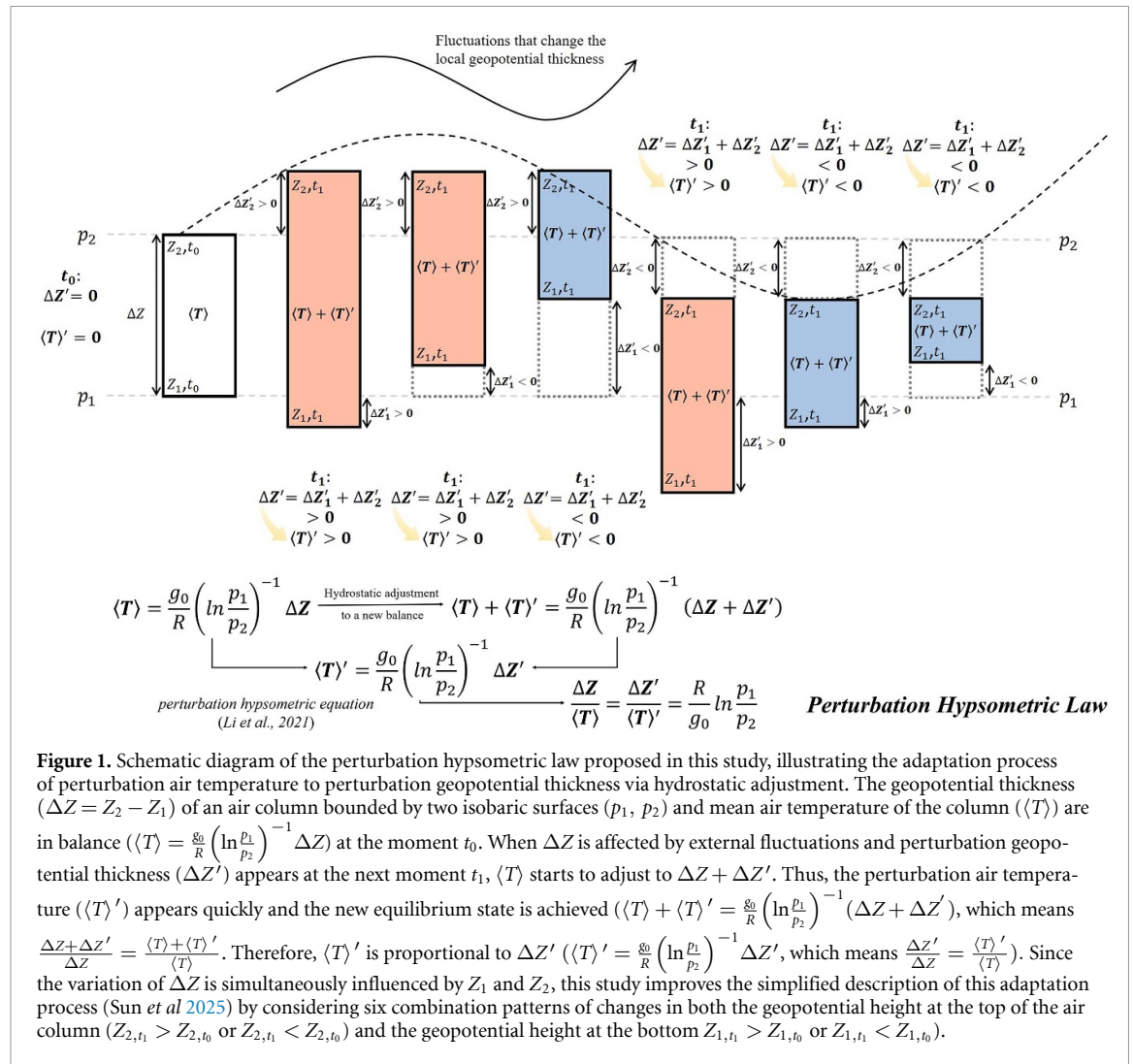
The perturbation hypsometric equation (Li *et al* 2022) relates perturbations in geopotential thickness to perturbations in layer-mean temperature:

$$\langle T \rangle' = \frac{g_0}{R} \left(\ln \frac{p_1}{p_2} \right)^{-1} \Delta Z', \quad (5)$$

where $\langle T \rangle'$ and $\Delta Z'$ represent the perturbations (i.e. departures from the equilibrium state at t_0) of $\langle T \rangle$ and ΔZ , respectively. This equation indicates that $\Delta Z'$ is proportional to $\langle T \rangle'$. Therefore, the proportionality constant between the two perturbation terms is identical to that derived from the initial hydrostatic equilibrium state before the external forcing is applied. The perturbation hypsometric law is mathematically expressed by the ratio form of the equilibrium and perturbation relationships:

$$\frac{\Delta Z}{\langle T \rangle} = \frac{\Delta Z'}{\langle T \rangle'} = \frac{R}{g_0} \ln \frac{p_1}{p_2}, \quad (6)$$

This law states that, for an air column bounded by two fixed isobaric surfaces, the ratio of geopotential thickness to layer-mean temperature is the same in the basic and perturbed hydrostatic states. It describes a linear constraint inherent to hydrostatic adjustment and is analogous to Gay-Lussac's law under the hydrostatic approximation. The applicability of



this relationship is not intrinsically region dependent. In principle, it can be applied to any atmospheric layer for which hydrostatic balance and the ideal-gas approximation are reasonably satisfied and that is bounded by fixed pressure surfaces. In this study, this law provides the theoretical basis for interpreting $\Delta Z'$ as a local bridge between geopotential height anomalies and surface cold extremes. Importantly, these six configurations demonstrate that a local temperature anomaly cannot be inferred from the geopotential height anomaly at a single pressure level. According to the perturbation hypsometric law, the layer-mean temperature anomaly is governed by the $\Delta Z'$, which is jointly determined by the geopotential height changes at the upper and lower boundaries of the air column. Thus, similar geopotential height anomalies at one pressure level may correspond to different temperature anomalies depending on the change at the other boundary. More details are provided in text S5. In this study, the perturbation hypsometric law is applied to the five key CHTe regions to theoretically interpret the relationship between ΔZ and local SATa. However, the applicability of this law is not confined to these five regions; rather, they

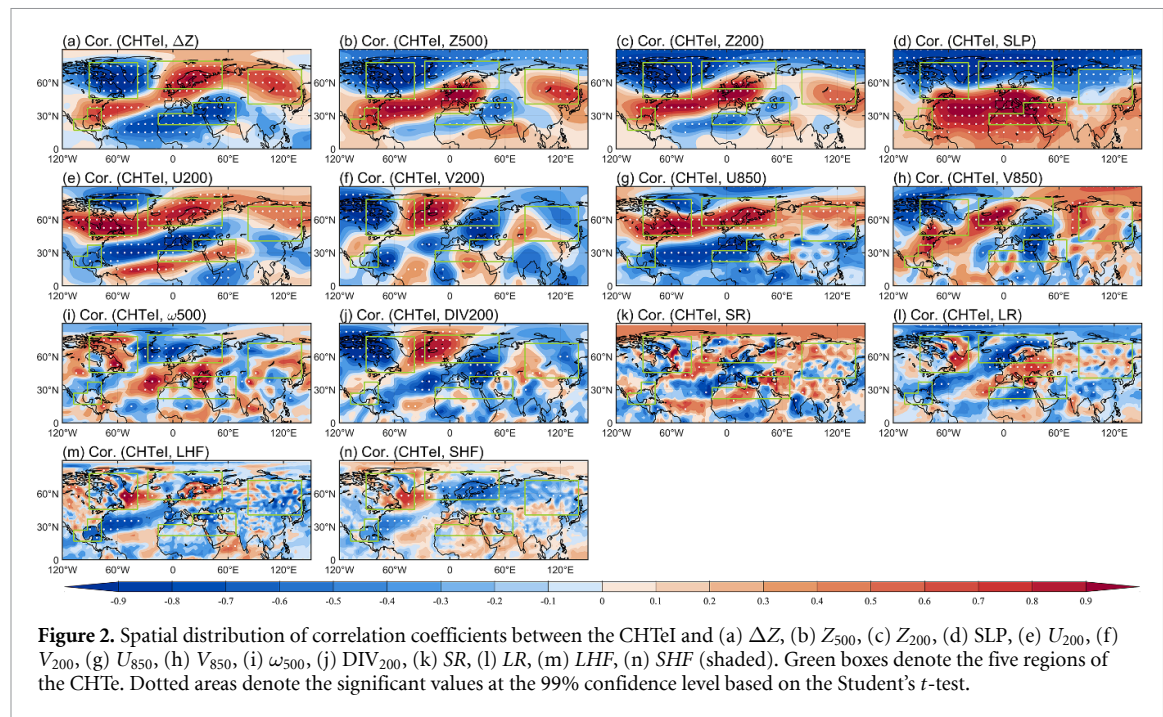
represent the specific regions examined in the present study.

To diagnose the propagation pathway, direction, and intensity of CHT-related Rossby wave activity, we use the horizontal Rossby wave ray flux, hereafter referred to as the Li–Yang WRF method (Yang and Li 2025). Details of the method are provided in text S6.

3. Results

3.1. Spatial linkages between the CHT and local atmospheric factors

The NAT can excite a Rossby wave train associated with the atmospheric teleconnection pattern CHT, which propagates downstream and modulates atmospheric-thickness anomalies over the five regions of CHTe, producing a corresponding ‘positive–negative–positive–negative–positive’ five-center structure (Wang et al 2025). Against this large-scale background, the present study further focuses on identifying which local atmospheric factors are most closely associated with the occurrence of ECEs



after the large-scale teleconnection signal reaches each key region.

Consistent with this large-scale background, the Li–Yang WRF diagnosis indicates that Rossby wave activity originating from the central North Atlantic propagates eastward along a pathway broadly corresponding to the CHT regions (blue boxes in figure S2), which are distinct from the five CHTe regions used to define the CHTel. This preferred pathway passes through four of the five CHT centers but does not directly intersect the Greenland center (text S7). This diagnosis provides dynamical support for interpreting the CHT as an organized Rossby wave pathway transmitting North Atlantic circulation anomalies downstream.

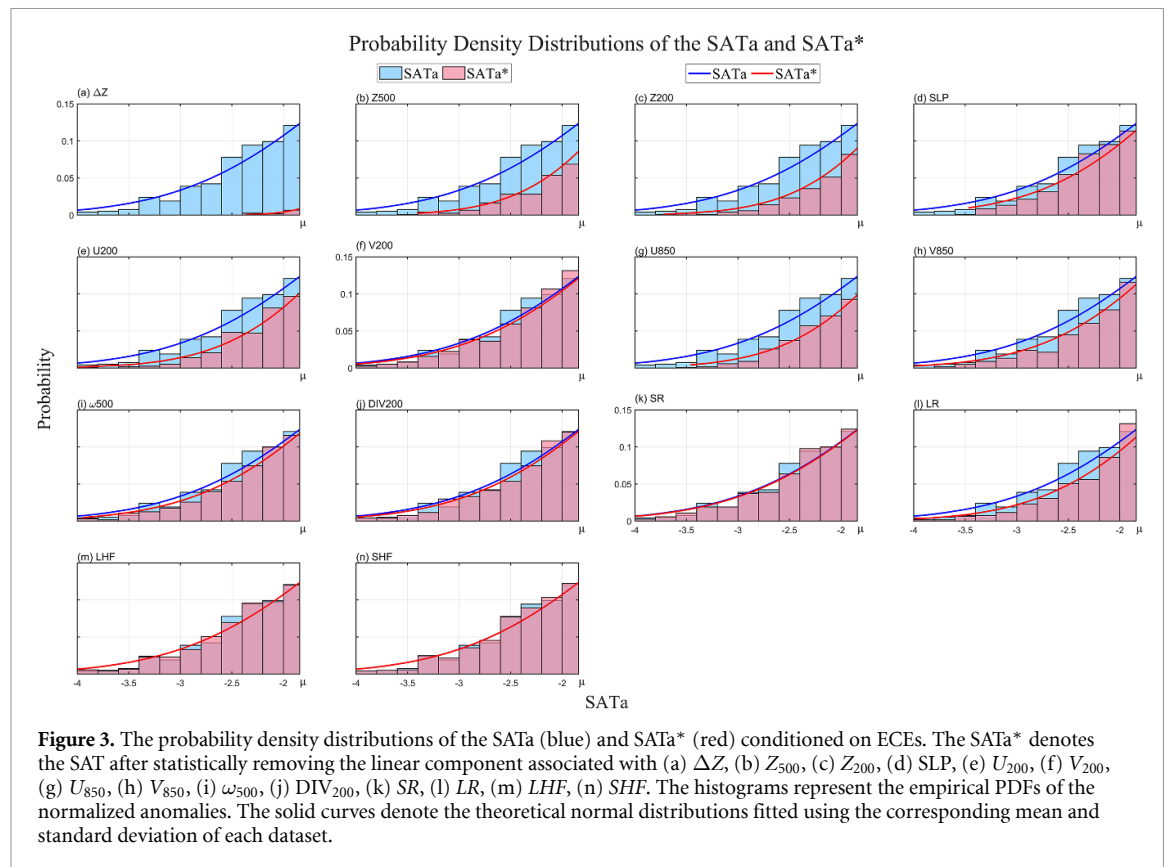
We then examine which local atmospheric factor is most closely associated with the hemispheric circulation pattern and may therefore provide a potential bridge to regional ECEs. All 14 local factors exhibit varying degrees of spatial association with the CHTel (figure 2), but their spatial coherence and regional consistency differ substantially. Several factors show pronounced correlations in individual regions, whereas these signals are relatively localized or vary in sign among the five CHTe regions. In contrast, ΔZ exhibits a more spatially coherent relationship with the CHTel, with correlation signs that systematically correspond to the alternating circulation anomalies of the five CHTe centers. This contrast suggests that ΔZ may better capture the five-region structure of the CHTe than the other local atmospheric factors. We therefore quantitatively evaluate their five-region spatial coherence using the area-weighted regional-mean correlation and the fraction of significant area within each CHTe region (figure S4). First, we calculated

the area-weighted regional-mean correlation coefficients of the 14 local atmospheric factors within each of the five CHTe regions (figure S4 (a)). The results show that, in the order of Southeastern North America, the Baffin Bay Coast, Northern Europe, the Middle East–North Africa, and Eastern Siberia, only ΔZ exhibits a clear and continuous positive–negative–positive–negative–positive alternating correlation pattern, showing a coherent spatial correspondence with the five-region structure of the CHTe. In contrast, the other local atmospheric factors do not exhibit an equally consistent alternating pattern across all five regions. We further calculated the fraction of regional area showing significant correlations within each of the five regions (figure S4 (b)). ΔZ exhibits consistently large significant-area fractions across all five regions, whereas the other local atmospheric factors generally show substantially smaller significant-area fractions in one or more regions. Taken together, the spatial sign structure of the regional-mean correlations and the significant-area fractions indicate that, among the 14 local atmospheric factors examined, ΔZ exhibits the most consistent and spatially coherent five-region statistical correspondence with the CHTel.

3.2. Assessment of local factors on ECEs

To quantitatively compare the associations of the candidate local factors with ECEs, we employ five complementary statistical metrics in the following analyses.

The PDF analysis provides statistical evidence for associations between local atmospheric factors and ECE occurrence. The rationale is that if a local factor has a strong statistical relationship with ECE occurrence, statistically removing its linearly associated



temperature component should produce a relatively large change in the ECE-related cold-tail distribution. As shown in figure 3, after removing ΔZ , the cold-tail probability density of SATa*, representing ECEs, decreases markedly relative to that of SATa, with $\Delta PDF < 0$. In contrast, removing any of the other 13 factors produces much smaller changes in the cold-tail probability density, with $\Delta PDF = 0$. This result indicates that removing ΔZ produces the largest change in ECE-related cold-tail probability among the factors examined. Paired contingency analysis and McNemar's test further support this conclusion (table S5). For example, as shown in table S6, removing ΔZ reduces the number of days classified as ECEs by 98%. To quantify cold-tail distributional differences without dilution by the bulk distribution, we further compute the tail K–L and J–S divergences; larger values indicate greater differences within the selected tail region. As shown in figure S5, after ΔZ is removed, the tail K–L divergence reaches 3.68, more than thirty times the second-ranked value for U_{850} (0.12); the tail J–S divergence for ΔZ is 0.14, whereas the values for all other factors do not exceed 0.02. Together, the qualitative PDF changes and quantitative tail-divergence metrics demonstrate that removing ΔZ substantially alters the extreme cold-tail distribution, whereas removing the other factors has little effect. This further indicates that ΔZ shows the strongest association with ECE occurrence among the local factors examined.

The conditional probability $P(\text{ECEs} \mid \text{extreme factor})$ quantifies the likelihood of ECE occurrence when a local atmospheric factor is in its extreme state. As shown in figure S6, $P(\text{ECEs} \mid \text{extreme } \Delta Z)$ reaches 65%, whereas the second-highest value, for Z_{200} , is only 37%, and all remaining factors are below 32%. This contrast indicates that extreme ΔZ is more strongly linked to ECE occurrence than the extreme states of the other local atmospheric factors. This result complements the PDF-removal evidence: the PDF analysis asks whether removing a factor changes the cold-tail distribution, whereas the conditional probability asks whether the extreme state of that factor preferentially occurs with ECEs. The agreement between these two independent perspectives strengthens the interpretation that extreme atmospheric thickness anomalies are closely tied to ECE occurrence.

We further evaluate the spatiotemporal coincidence between extreme local atmospheric factors and ECEs using the extreme consistency rate. From 10 d before the ECEs to Day 0 (figure S7), the consistency rate between extreme atmospheric thickness and ECEs is generally higher than those associated with the other extreme local atmospheric factors, indicating a more pronounced preceding statistical correspondence between extreme atmospheric thickness and ECEs over this period. For extreme atmospheric thickness itself, however, the consistency rate reaches its maximum at Day 0 and gradually decreases with

increasing lead time. When atmospheric thickness leads the ECEs by 5 d, the consistency rate falls below 40%, indicating that the statistical correspondence between extreme ΔZ and ECEs is mainly concentrated within a lead time of approximately 0–4 d.

To complement the daily-scale analyses above, we conduct an additional annual-scale analysis using 2×2 contingency tables. Among the 7 years when extreme ΔZ occurs, 5 coincide with ECEs, and 2 do not; in contrast, among the 60 years with non-extreme ΔZ , only 2 experience ECEs while 58 do not (table S7). The corresponding χ^2 value for ΔZ is 31.07, significant at the 99.9% confidence level, and $\varphi = 0.68$, approximately twice the φ value of the next-ranked factors and much larger than those of most other variables (table S8). This demonstrates a strong annual-scale association between extreme ΔZ and ECE occurrence, further supporting ΔZ as the factor exhibiting the strongest long-term statistical association with ECEs among those examined. The contingency analysis also indicates that the ΔZ –ECE relationship is strong but not deterministic. Although ECEs are much less frequent in non-extreme- ΔZ years, some ECE years occur without ΔZ crossing its prescribed extreme threshold, indicating that extreme ΔZ is not a strict necessary or sufficient condition for ECE occurrence. This asymmetry is physically consistent with ΔZ representing the layer-mean thermal state of the 1000–500 hPa atmospheric column, whereas ECEs are defined by near-surface minimum temperature and can additionally be modulated by shallow near-surface and local dynamical and thermodynamical processes.

In linear-correlation analysis, ΔZ shows the strongest correlation with ECEs ($r = 0.83$, 68.5% explained variance), followed by Z_{200} ($r = 0.46$), Z_{500} ($r = 0.46$), and V_{850} ($r = 0.44$; table S9). The ΔZ –ECE association remains significant after statistically controlling for the other leading factors, whereas the associations of these factors with ECEs become weak and nonsignificant after adjustment for ΔZ (table S10). This indicates that ΔZ retains a robust statistical relationship with ECEs beyond the shared variance captured by the other leading factors. Adding the other leading factors provides only limited additional explanatory power once ΔZ is included, while the reduced associations of these factors after adjustment for ΔZ indicate substantial shared variance with ΔZ . This result is physically consistent with the perturbation hypsometric law, which provides a hydrostatic interpretation of the close relationship between ΔZ and temperature. Overall, these analyses further identify ΔZ as the factor exhibiting the most robust statistical correspondence with ECEs among the variables examined (figure S8).

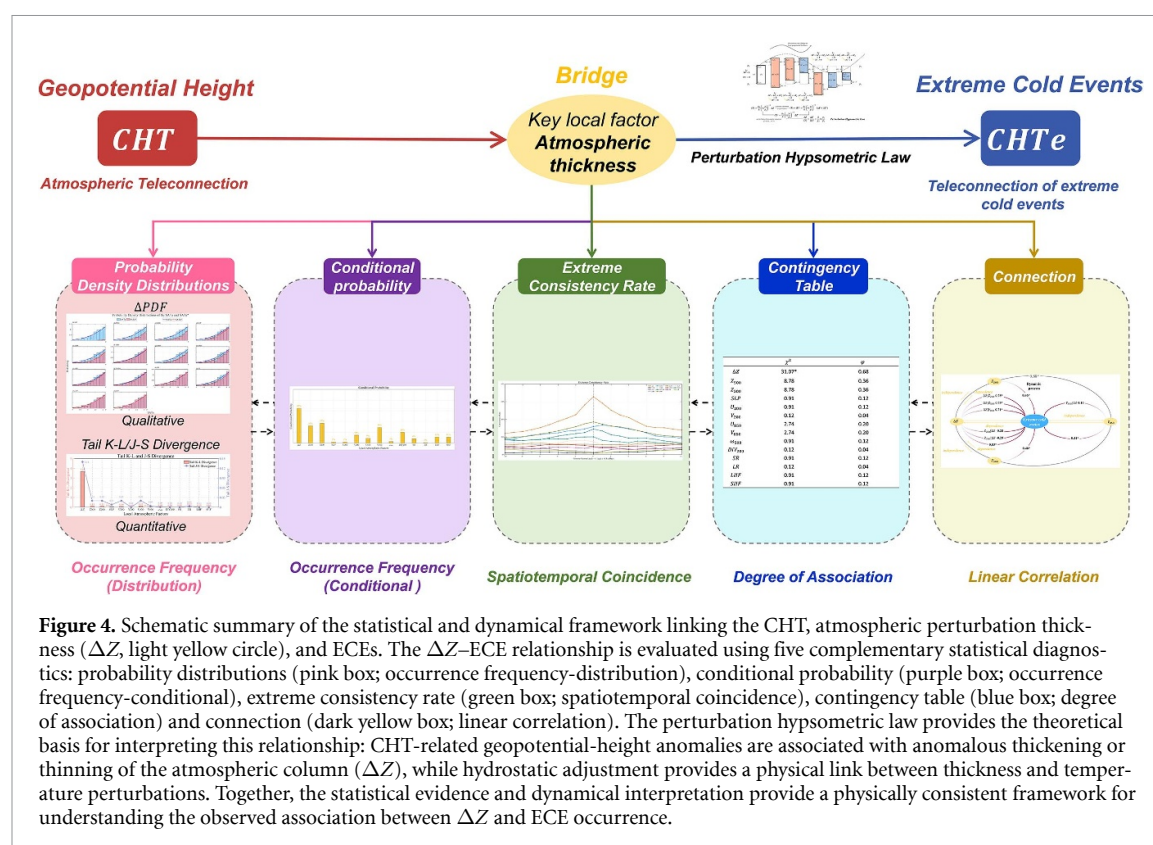
4. Conclusions and discussion

In this study, atmospheric thickness between 1000 and 500 hPa (ΔZ) exhibits the strongest and most robust statistical association with regional ECEs among the 14 local atmospheric factors examined. To provide a physical interpretation of this statistical relationship, we propose the perturbation hypsometric law, which states that perturbations in geopotential thickness and layer-mean temperature share the same hydrostatic proportionality as their basic-state counterparts. This law provides the physical basis for the atmospheric-thickness pathway: CHT-related geopotential-height anomalies can alter the thickness of the 1000–500 hPa air column, and hydrostatic adjustment links this thickness perturbation to a corresponding layer-mean temperature perturbation. Thus, negative ΔZ anomalies indicate a thinner and colder lower-to-middle tropospheric column, favoring surface cold extremes.

The multidimensional statistical framework further shows that ΔZ is not only spatially coherent with the CHTe pattern, but also consistently exhibits the strongest association across complementary statistical diagnostics (figure 4). Removing ΔZ produces the largest change in the extreme cold-tail PDF and the largest tail-divergence values. Extreme ΔZ also yields the highest conditional probability of ECE occurrence and the strongest lead-lag consistency at 0–4 d, suggesting useful short-lead information. Annual contingency analysis further supports a strong long-term association, while partial-correlation and multiple-correlation analyses show that adding Z_{500} , Z_{200} , or V_{850} provides little additional explanatory power once ΔZ is included. Together, these results indicate that atmospheric thickness is a measurable and physically interpretable bridge from hemispheric teleconnection dynamics to regional winter cold extremes.

The temperature tendency equation and composite evolution analysis further provide dynamical context for interpreting the physical processes underlying the ΔZ –ECE relationship (text S8).

Across the five CHTe regions, the ΔZ –ECE relationship remains consistently robust, whereas the secondary associations show clear regional dependence (text S9). The primary datasets used in this study are at daily resolution; therefore, the lead-lag relationships identified here are confined to timescales of several days. When extreme atmospheric thickness lags the ECEs, the consistency rate is generally lower than when atmospheric thickness leads the ECEs (figure S7). Nevertheless, the perturbation hypsometric relationship between atmospheric thickness and surface air temperature is not restricted to daily data and is also applicable to monthly and annual datasets (e.g. Wang *et al* 2024, 2025, Sun *et al* 2025), suggesting its potential relevance for subseasonal and climate-scale



prediction. The lead–lag relationship between atmospheric thickness and surface air temperature warrants further investigation across different timescales.

Several limitations and future directions deserve mention. The statistical framework is primarily linear and does not establish causality or fully account for nonlinear interactions. We cannot rule out the possibility that other local factors may exert synergistic effects (Li *et al* 2019) on ECEs. The perturbation hypsometric law identifies the hydrostatic constraint linking thickness and temperature, but targeted numerical experiments are still needed. Finally, extreme events often arise from complex interactions across multiple temporal and spatial scales, involving a range of factors and physical processes. Accurately identifying these factors and quantifying their contributions will require the development of novel methodologies and theories.

Acknowledgments

Thanks for the helpful discussions and suggestions from Dr Hui-Huang Fu. This work is jointly supported by the National Natural Science Foundation of China (NSFC) Project (42288101), Laoshan Laboratory (No. LSKJ202202600), Fundamental Research Funds for the Central Universities (202242001, 202515028, 202661006). We extend our sincere gratitude to the NCEP-NCAR for providing the reanalysis data, and the Met Office Hadley Centre for providing the HadEX3 dataset.

Data availability statement

The atmospheric datasets used in this study include: The NCEP/NCAR reanalysis data set is available at <https://psl.noaa.gov/data/gridded/data.ncep.reanalysis.html>. The TN10p data is available at www.metoffice.gov.uk/hadobs/hadex3/download.html.

Supplementary data available at <https://doi.org/10.1088/1748-9326/aea2b0/data1>.

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