

Cold spells induced by slow and amplified atmospheric waves

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Abstract. Cold spells in the Northern Hemisphere mid-latitudes have been linked to Rossby waves. Yet the mechanisms by which these waves impact cold-spell formation remain unclear. Here we develop novel metrics to separately measure the amplitude and speed of ridges and troughs, examining the behavior of Rossby waves during winter cold spells. Our findings indicate that while ridges and troughs across the entire mid-latitudes experience significant changes during cold spells, the local ridge and trough near the cold spell's location play a major role in the development of these events. The nearest upstream ridge and downstream trough of the cold-spell region are located in a way that facilitates development of the extreme cold anomaly. This ridge and trough amplify and slow, enhancing and prolonging southward advection of cold air from the Arctic into the cold-spell region. The slow and amplified upstream ridge and downstream trough occur several days before the region's minimum temperature, suggesting these local wave anomalies induce cold-spell formation.

10 1 Introduction

High-amplitude quasi-stationary Rossby waves may generate extreme weather events (e.g., Hoskins and Woollings, 2015; Fragkoulidis and Wirth, 2020; White et al., 2022), and increase in temperature extremes in the mid-latitudes is argued to be linked to increased upper-atmosphere waviness (e.g., Fragkoulidis et al., 2018). An increase in meridional amplitude of Rossby waves may enhance the meridional exchange of warm and cold air, leading to the advection of cold air to the south and warm air to the north (Zschenderlein et al., 2018; Jolly et al., 2021). Some studies link extreme temperatures to a significant amplified circumglobal waviness based on wave amplitudes of individual wave numbers (Petoukhov et al., 2013; Screen and Simmonds, 2014), while others assert a robust correlation between local waviness and temperature extremes (Röthlisberger et al., 2016; Fragkoulidis et al., 2018; Fragkoulidis and Wirth, 2020). Studies based on the amplified circumglobal waviness argued that free-traveling synoptic waves of some specific wave numbers (6, 7, or 8) may be trapped within the midlatitude waveguide leading to extreme temperature events (Petoukhov et al., 2013; Kornhuber et al., 2017), whereas others caution against overemphasizing the significance of circumglobal waviness, suggesting that local climatic factors may play a more critical role in influencing temperature extremes (Teng and Branstator, 2019).

Slow movement of Rossby waves may also lead to weather persistence, for instance in terms of blocking, which may develop into extreme weather events, such as floods, droughts, heat waves, and cold spells (Francis and Vavrus, 2012; Pfliederer and Coumou, 2018; Riboldi et al., 2020; Jolly et al., 2021; Wicker et al., 2024). Extreme rainfall over Eastern Australia has been linked to slow-moving upper-level low-pressure systems (Barnes et al., 2023; Reid et al., 2025; Vries et al., 2025), and it is

argued that slow propagation of atmospheric blocking, identified using a blocking cell-tracking algorithm, lead to significant surface temperature anomalies (van Mourik et al., 2025). The propagation speed of atmospheric blocking can impact the upstream and downstream of the block, in the way that rapidly westward-moving blocks lead to less persistent cold events in the upstream of the block, while slower westward-moving or quasi-stationary blocks lead to strong cold anomalies in the downstream of the block (Chen and Luo, 2017; Yao et al., 2017).

The frequency of extreme events has risen in recent decades due to anthropogenic warming (IPCC; Seneviratne et al., 2021), and these events are likely to become more intense and break the previous extreme records by a large margin in the following decades (Fischer et al., 2021). In addition, despite the warming Earth, cold spells in some regions of the Northern Hemisphere have recently increased (e.g., Cohen et al., 2021, 2024), although, in general, cold spells are projected to become less frequent by the end of the century (IPCC; Seneviratne et al., 2021), and the likelihood of experiencing the strongest historical extreme cold spell events is expected to diminish (Ribes et al., 2025).

Increases in extreme event frequency, specifically cold spells, have been linked to Arctic amplification (AA), a phenomenon by which the Arctic warms faster than the mid-latitudes (e.g., Francis and Vavrus, 2012; Cohen et al., 2014): the AA is weakening the jet stream, making the large-scale atmospheric waves more wavy (Francis and Vavrus, 2012; Coumou et al., 2015; Stendel et al., 2021); in addition, AA causes Rossby waves to become more stationary, resulting in an increase in the frequency of extreme weather (Francis and Vavrus, 2012). This hypothesis, however, has been questioned by several studies (e.g., Barnes, 2013; Screen and Simmonds, 2013; Blackport et al., 2022, 2024).

The dispute about the influence of the AA on planetary waves might stem from the sensitivity of waviness assessments to the applied metric (Geen et al., 2023), highlighting the need for improved and universal wave metric. Existing wave amplitude metrics, such as Rossby wave packets (e.g., Fragkoulidis et al., 2018; Röthlisberger et al., 2019), local jet waviness (e.g., Röthlisberger et al., 2016), and hemispheric variability (e.g., Petoukhov et al., 2013; Kornhuber et al., 2019), do not establish the role of the amplitude of each ridge and trough in temperature events. In addition, as far as the authors know, the speed of local ridges and troughs and their role in extreme events have not earlier been quantified. Some studies estimate atmospheric phase propagation by following grid points using lag-correlation in different bandpass-filtered geopotential height maps, tracking the maximum positive correlation center (e.g., Blackmon et al., 1984; Takaya and Nakamura, 2001). Others derive phase-speed spectra for each zonal wavenumber by performing a space–time spectral decomposition of upper-tropospheric winds along latitude circles (e.g., Randel and Held, 1991; Domeisen et al., 2018; Riboldi et al., 2020). A local approach is also used to estimate the speed of Rossby wave packets (e.g., Fragkoulidis and Wirth, 2020; Fragkoulidis, 2022). Here, we aim to explore the linkage between cold spells and the amplitude and speed of each ridge and trough. Cold spells are particularly of interest in the context of Rossby waves because these cold extremes are argued to be primarily driven by the large-scale advection of cold air from higher latitudes (Bieli et al., 2015; Tuel and Martius, 2024). Hence, this study further investigates the role of individual ridges and troughs in causing cold spells.

2 Methods

60 In this study, we used the European Centre for Medium-Range Weather Forecasts (ECMWF) analysis 5th Generation (ERA5) dataset with a horizontal resolution of $0.25^{\circ} \times 0.25^{\circ}$ (Hersbach et al., 2020) for the period 1978-2024. The study is for the winter season (December-January-February) and the satellite era. Minimum daily temperature and daily average temperature are obtained based on hourly ERA5 data. Additionally, 3-hourly geopotential height data at 300 hPa, 500 hPa, and 850 hPa are utilized.

65 2.1 Cold spells

To identify cold spells and their magnitudes, the cold wave magnitude index daily (CWMId; Morlot et al., 2023) is applied at each horizontal grid point. The daily cold spells index is based on at least three consecutive days below a daily threshold, defined as the 90th percentile of anomalies relative to the climatology of minimum daily air temperature smoothed with a 31-day running mean filter. The climatology is based on the reference period 1981–2020. The background trend is small relative to the daily anomalies and we have checked that it does not affect the overall conclusions drawn in the paper (not shown).

70 2.2 Study regions selection

To select cold spell regions, we first identified historical extreme cold days on the hemispheric scale. To do so, a non-standardized version of the mid-latitude extreme (MEX; Riboldi et al., 2020) index is applied for land in four latitude bands (35-45°N, 45-55°N, 55-65°N, and entire mid-latitudes: 35-65°N). Dividing the mid-latitudes into three sub-bands enables an identification of extreme cold days in the lower, middle, and upper sections of the mid-latitudes. The MEX provides an overall measure of mid-latitude temperature daily variance by averaging over each grid point in the mid-latitudes the squared anomaly of the temperature. Here, "non-standardized" means the MEX index is used without subtracting its time-averaged mean and dividing by its standard deviation, as done in Coumou et al. (2014). The top 1% of daily MEX values—a criteria to classify cold extreme days—within all the winters during the period 1978-2024 yields 125, 117, 119, and 141 days for the bands 35-45°N, 45-55°N, 55-65°N, and 35-65°N, respectively. To identify the areas providing the largest cold anomaly contribution to the top 1% MEX days of each latitude band, the CWMId (Fig. 1) and cold anomaly (supplementary Fig. S1) of the extreme cold days are calculated at each grid point in the mid-latitudes. Since these two diagnostics yield similar results, we focus on CWMId in the analysis.

Fig. 1 shows the composite of the geopotential height (gph) anomaly at 500 hPa and the cold-wave magnitude index for cold extreme days for each latitude band. For the top 1% MEX days over the whole midlatitude (35-65°N; Fig. 1a), cold spells are more intense in the northwestern part of North America and northwestern Eurasia. The intensity decreases toward the south and east in both continents. The gph anomaly is positive over Greenland and the North Pacific Ocean, the two primary blocking regions (Barriopedro et al., 2006). Over the North Atlantic Ocean, a negative gph anomaly is encountered that may indicate the negative phase of the North Atlantic Oscillation (NAO) during these cold events, a phenomenon that promotes European winter cold spells (e.g., Hanna and Cropper, 2017; Vihma et al., 2020).

For the top 1 % MEX days over 35-45°N, little or no positive gph anomaly is encountered over Greenland, which is consistently associated with a weak intensity of CWMId over Europe (Fig. 1b). The main gph anomaly consists of a positive height anomaly over Alaska and a negative height anomaly downstream over West North America. Consistently, in this latitude band, CWMId is more pronounced over northwest North America, the central US, and Central Asia. Intense CWMId over Central Asia appears to differ from that of the central US, as it is likely not linked with higher latitudes, leading us to speculate that the cold spells in this region do not originate from Arctic cold air. Therefore, this region is not included in our study.

For the top 1 % MEX days over 45-55°N, in contrast to 35-45°N, there is a strong positive gph anomaly upstream of Europe, centered over Iceland, while the positive gph anomaly is weak upstream of North America (Fig. 1c). For this latitude band, the CWMId is strong over the middle and east of Europe and over the west of Canada. For the top 1 % MEX days over 55-65°N, CWMId is stronger at higher latitudes, encompassing Scandinavia, the west of Canada, and Northern Siberia (Fig. 1d). For this latitude band, there are two strong positive gph anomaly centers over Baffin Bay and the Bering Sea. The top 1 % MEX over the 45-55°N and 55-65°N, similar to MEX across the entire midlatitude, exhibits a negative gph anomaly over the North Atlantic Ocean.

Based on the results of the composite of days with the top 1 % MEX over these four latitude bands, 9 regions are selected for the study of the impact of large-scale upper atmospheric circulation on cold spells, shown as hatched areas in Fig. 2. The regions include Northwest Canada (R01), Northwest US and Southwest Canada (R02), Central US (R03), British Isles (R04), Scandinavia (R05), Europe (R06), Northwest Russia (R07), Southwest Russia (R08), and Northern Siberia (R09). If 50 % or more of the grid points within each region experience cold spells on a given day, it is considered a cold spell day in that region. Applying grid-point thresholds between 20 percent and 70 percent are also tested to ensure that the drawn overall conclusions remain unchanged for an altered threshold. The number of days with 50% or more of the grid points experiencing cold spells in each region, as based on the CWMId, is 239, 184, 164, 157, 170, 171, 195, 150, and 152, respectively, for R01 to R09.

2.3 Significant test

To test the null hypothesis that anomalies associated with the cold spells are zero, a Monte Carlo simulation with random samples is employed. Specifically, the average number of cold spell days across each region is compared with the average of an equivalent number of randomly selected days. Instead of selecting random days from the entire winter season, the selection is weighted based on the frequency distribution through the season of cold spell days, since cold spells appear to occur more frequently in some parts of winter. This Monte Carlo test is repeated at least 2000 times to ensure a sufficient sample size, and significant differences from zero are identified at the 95% confidence level.

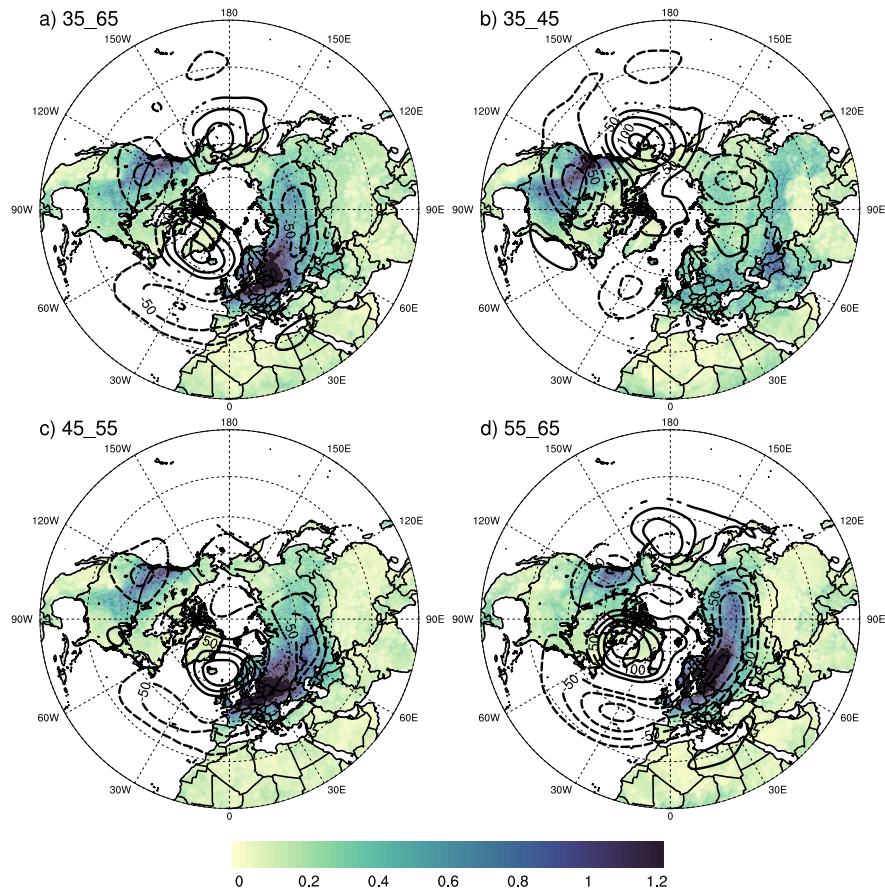


Figure 1. The composite of the geopotential height anomaly relative to climatology (m) at 500 hPa (lines; only significant at the 95% confidence level) and the cold wave magnitude index daily (CWMID; colors) for cold extreme days given by the top 1 % of the MEX over land at the latitude band of a) 35–65°N, b) 35–45°N, c) 45–55°N, and d) 55–65°N.

3 Wave metrics

120 3.1 Trough and ridge speed

In order to investigate the role of large-scale waves, a Fourier decomposition of gph is employed to separate the field into distinct harmonic waves based on the zonal wave number. The combination of the first five waves is regarded as an estimation of large-scale waves (Z1–5; solid line in Fig. 3a); these are planetary or Rossby waves (except possibly at high latitudes, where it may be argued that their wave length are too short for the categorization as planetary waves). We estimate the speed of atmospheric wave zonal propagation by utilizing a ridge and trough tracking algorithm (Fig. 3a). Small amplitude ridges and troughs are ignored since they are often short lived (red stars in Fig. 3a). This is accomplished by considering a threshold for the gph at each local maximum (ridge) and local minimum (trough): At each time step (3-hourly) and latitude, the gph at the ridge

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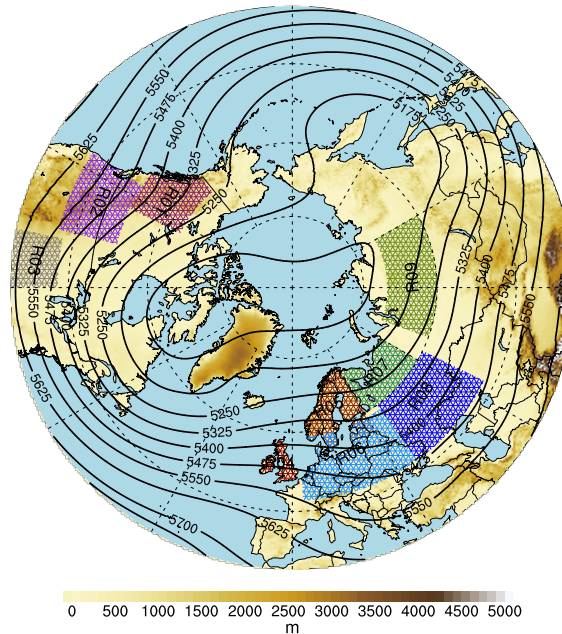


Figure 2. The chosen study regions (hatches). The winter climatology mean geopotential height field (m) and the elevation above mean sea level (m) are shown by contours and shading, respectively.

(trough) position must be greater (less) than the gph at 30° to the west and east of the ridge (trough) position by at least 8 % of the Z1–5 amplitude (see blue texts and arrows in Fig. 3a). The Z1–5 amplitude is defined as the difference between the highest
130 ridge and the lowest trough at all longitudes around a latitude circle at each time step and latitude. These criteria were carefully chosen after multiple experiments to ensure that all large waves are tracked while minor waves are neglected. Nonetheless, in all the experiments, as long as the thresholds do not suppress large waves, the drawn overall conclusions regarding wave speed changes remain unchanged for other parameter settings. We track each local ridge and trough in time over longitudes to find the ridge and trough speed, respectively. Also, we save all local ridge latitude and longitude (RLL) and trough (TLL) positions.
135 At each latitude, ridges and troughs are labeled according to their positions (longitude) and are tracked to the next time step by searching for the nearest ridge or trough within a 12-degree limit both to the west and east of their location (Fig. 3b). If no corresponding feature is detected within this range, the ridge or trough is considered to have decayed. Conversely, if a new ridge or trough appears without a corresponding feature in the previous time step, it is considered to have formed.

Note that the ridge and trough speeds as defined here are not the same as the phase or the group speed. Phase speed represents the speed of individual wave numbers, whereas the group velocity is the propagation of the wave «envelope». These are both different from the trough and ridge velocity of combined waves as investigated here. Consequently, here we will refer to the examined velocity simply as the wave speed or the speed of ridges and troughs, intentionally avoiding the terms phase speed or group speed.

Fig. 4 displays the probability of the winter climatological positions of ridges and troughs, as well as the decay and formation of these features. Most often three major ridges and three major troughs can be seen throughout winters (Fig. 4 and supplementary Fig. S4). The three main ridges are located over the west coast of North America (hereafter, WNA ridge), the western flank of the Tibetan Plateau (hereafter, TP ridge), and the North Atlantic Ocean (hereafter, NAO ridge, at low latitude over the middle part, and at high latitude over the eastern part of the ocean). The three main troughs are located over Eastern North America (hereafter, ENA trough), the eastern Mediterranean (hereafter, EMed trough), and East Asia (hereafter, EAsia trough). The daily speed of these six major ridges and troughs is determined by summing their 3-hourly values across their respective longitudinal ranges (Fig. 3c). The speeds of the ridges over 180°W – 90°W , 60°W – 30°E , and 30°E – 120°E are added to calculate the daily speeds of the WNA ridge, the NAO ridge, and the TP ridge, respectively. The daily speeds of the ENA trough, EMed trough, and EAsia trough are also found by taking the sum of the speeds of the troughs over 120°W – 30°W , 0°E – 90°E , and 90°E – 180°E , respectively (Fig. 3c).

Ridges and troughs typically form upstream of their climatological positions and decay downstream of them (Fig. 4). The WNA ridge, NAO ridge, and ENA trough exhibit stronger formation and decay at high latitudes compared to mid-latitudes, indicating that these ridges and troughs tend to persist longer in the midlatitudes. The TP ridge and EMed trough show large formation and decay at mid-latitudes, likely due to their interaction with the complex orography. Nonetheless, the likelihood of their decay and formation remains significantly lower than their climatological existence, indicating persistency of the ridges and troughs (Fig. 4). The EAsia trough forms more strongly at midlatitudes, while its decay is more pronounced at higher latitudes. Over the Pacific Ocean, there is one ridge and one trough formation center, both of which decay downstream from their formation center. As can be seen from the climatological position of ridges and troughs, there is no strong ridge or trough in these locations, suggesting that these features are short-lived.”

3.2 Trough and ridge amplitude

Several metrics have been developed to study the waviness of the atmosphere (e.g., Francis and Vavrus, 2012; Barnes, 2013; Hassanzadeh et al., 2014; Chen et al., 2015; Cattiaux et al., 2016; Geen et al., 2023), many of which rely on the meandering of the 500 hPa isohypses (lines of constant gph) as a proxy. Considering the variations of wave amplitude with different isohypses—e.g., associated with the seasonal variability and poleward shift due to global warming (Barnes, 2013)—selecting an appropriate isohypses are critical. To address these challenges, Cattiaux et al. (2016) derived isohypses from the daily average of the gph at 500 hPa between 30°N and 70°N , hereby varying in time. They regarded the meandering of these isohypses as an indicator of the waviness of the atmospheric flow at 50°N .

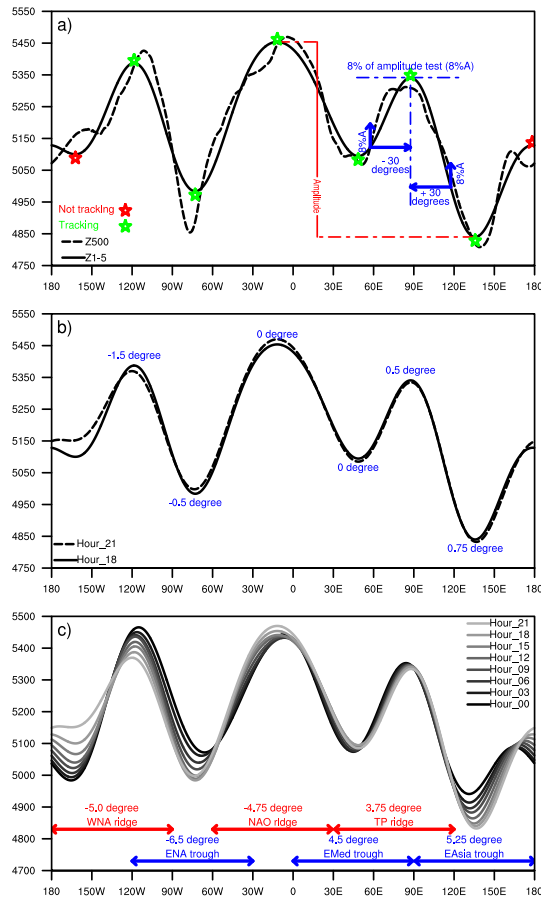


Figure 3. The 500 hPa geopotential height field (Z500; dashed, black line) and the sum of the first five zonal Fourier-decomposed waves of Z500 (Z1–5; solid, black line) for a randomly selected time step (January 7, 2003) at 60°N. The blue text and arrows are associated with a threshold metric that identifies the strength of the ridges and troughs (see method for more information). Based on this threshold, three ridges and three troughs (green stars) are high and deep enough, respectively, for being detected by the tracking algorithm. In this time step, the tracking algorithm will ignore one ridge and one trough (red stars) that do not meet this threshold. b) The Z1–5 for two consecutive time steps (3 hours) and the changes in the ridges' and troughs' longitudinal location between these time steps. c) The Z1–5 for one day (8 time steps) and the calculated longitudinal shifts of each ridge and trough over the entire day.

Here we focus on identifying changes in waviness at each latitude and at 300, 500, and 850 hPa, which requires some modification of the Cattiaux et al. (2016) approach: For a given study region, the most common isohypse at each time step and latitude is taken as the zonal average of the gph over half of the hemisphere centered on the center longitude of each study region. Then, the meridional wave extent for that latitude is defined as the meridional extent of the associated isohypse determined by calculating the difference between the maximum latitudinal location and the minimum location of that isohypse (total gph isohypse extent; TIE). By focusing on half of the hemisphere, it is ensured that at least one full ridge and trough are

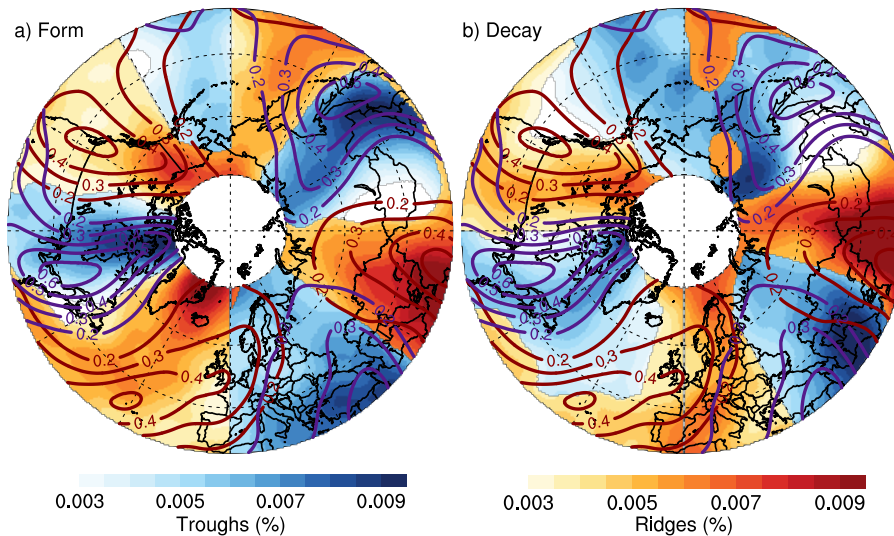


Figure 4. The probability of ridges and troughs and their formation and decay as a function of location. Shown with contours are the probability (in percentage) of 300 hPa ridges (brown) and troughs (purple) during winter climatology as a function of location. The red (blue) shading indicates the climatological probability a) formation and b) decay of ridges (troughs) as a function of location. To show both ridges and troughs in a single plot, priority is given to the stronger feature. A Gaussian function, using a weighted running average over eight degrees in latitude and over an equal distance in longitude, is used to smooth the probabilities.

included within the region. In a similar way, the meridional extent of the first five Fourier decomposition wave numbers of gph (Z1-5) is calculated (planetary isohypse extent; PIE). During cold spells in the defined regions, both TIE and PIE demonstrate the same significant pattern of changes (not shown). This shows the modest impact of small waves on the wave amplitude changes during cold spells. Therefore, we only consider the results of the meridional extent of Z1-5.

Some isohypses have unconnected subparts (see Fig. 5a), referred to as cut-off low and cut-off high, which might lead to interpretation errors in the wave amplitude. For instance, in Fig. 5a, at high latitudes (around 85°N), the most common isohypse is 8300 m, and at these latitudes, the waviness of this isohypse is small. But, due to a cut-off low of this isohypse at midlatitudes, the estimated wave amplitude at 85°N is large. Moreover, the PIE does not specify the separate contribution to

the wave amplitude from ridges and troughs. Understanding these contributions is critical, especially during cold spells, as it helps determine which parts of the wave—ridge or trough—having the greater impact.

Hence, we here propose an alternative metric where the meridional extent of the Z1-5 ridges and troughs are measured separately (Fig. 5b). The meridional extent of the ridge is given as the latitudinal difference between the RLL and the isohypse
190 determined by the zonal average of the Z1–5 over a 90-degree longitude range centered at the RLL (Fig. 5b, red line); hence, the meridional extent of a ridge at a given latitude is obtained as the difference between that latitude and the latitude of the first cross of the isohypse when moving northward along the longitude of the RLL. The employed method does not take into account the possible meridional tilting of waves. To address this, at 300 hPa, the wave amplitude is calculated for all longitudes within the 90-degree longitude range. The highest value obtained is then considered the meridional amplitude of the ridge.
195 While the highest value often yields slightly larger wave amplitudes than that obtained at the RLL, the overall conclusions regarding changes in wave amplitude during cold spells remain the same for both approaches (not shown). Therefore, to reduce computational usage, we used the amplitude at the ridge position. The meridional extent of troughs are obtained in a similar way but moving southward from TLL.

The estimated meridional amplitudes of a given ridge (trough) is highly influenced by the longitudinal range over which the
200 zonal average of the Z1–5 is taken for determining the isohypse. Using a smaller longitude range yields an isohypse value closer to the value of Z1-5 at RLL (TLL), which can lead to small amplitudes. Using a large longitude range might yield an isohypse that cannot be found near the RLL (TLL), resulting in a missing value for the amplitude. Therefore, after conducting multiple experiments, the 90-degree threshold is chosen. Nonetheless, despite the arbitrariness of the meridional wave amplitude when defined this way, its magnitude exhibits a consistent latitude-dependent pattern and comparable bands of significance in both
205 the climatology and cold-spell cases, which is important and sufficient for this study.

To compare the results of the refined method with PIE, considering the total wave amplitude, the average of the amplitude of the ridges plus those of the troughs over half of the hemisphere centered at each study region is calculated (ridge-trough isohypse extent; RTIE). For climatology, the RTIE amplitude (supplementary Fig. S2, gold lines) is nearly half of the PIE amplitude. During cold spells, there are likewise many similarities between the RTIE and the PIE, yet they have some important
210 differences. During cold spells, the PIE amplifies over a wider latitude range compared to the RTIE. This could be attributed to the influence of cut-off lows or cut-off highs on the amplitude of distant locations.

4 Results

4.1 Wave location

To investigate the impact of planetary waves on the formation of winter cold spells, nine regions with a high likelihood of
215 cold spells are selected (Fig. 2). These regions are chosen based on their significant contribution to the historical extreme cold days on a hemispheric scale (see methodology section). During cold spells over North America (R01, R02, and R03), large-scale gph anomaly patterns show an anticyclonic gph anomaly over the North Pacific and a cyclonic gph anomaly over North America (supplementary Fig. S3a, b, and c), similar to the findings of other studies (e.g., Xie et al., 2017). During cold

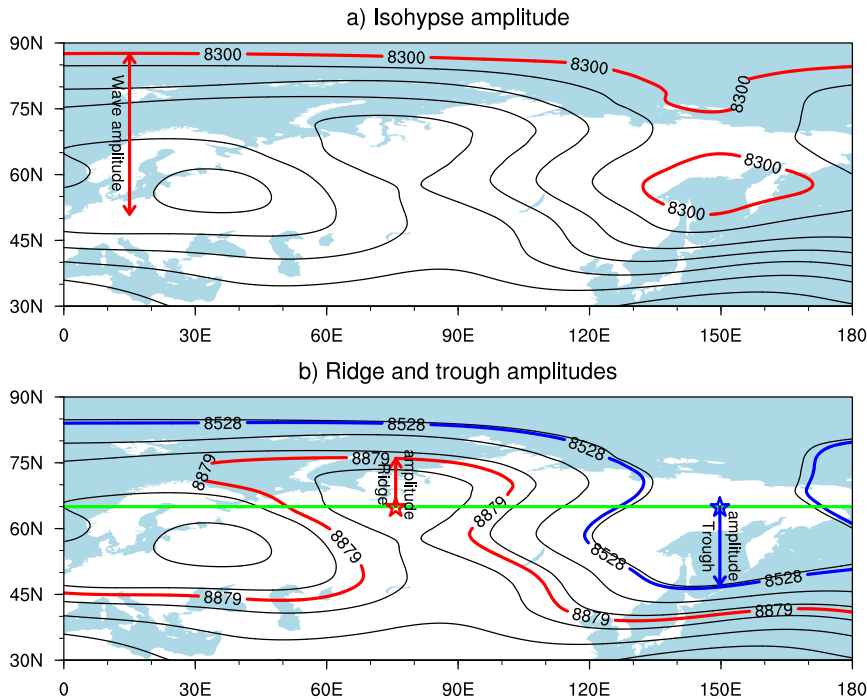


Figure 5. The sum of the first five zonal Fourier-decomposed waves of the 300 hPa geopotential height field (Z1-5) for a randomly selected 3-hourly time step (January 4, 2016, at 18:00). a) The 8300 isohypse and its amplitude are highlighted; b) The ridge (red arrow) and trough (blue arrow) meridional amplitudes at the ridge (red star) and trough (blue star) center at latitude 65°N (green line). The 8879 isohypse (red line) and the 8528 isohypse (blue line) are determined by the zonal average of the Z1-5 over a 90-degree longitude range centered at the ridge and trough positions, respectively, at 65°N.

spells, the locations of the WNA ridge and the ENA trough shift to the west of their corresponding climatological positions (Fig. 6a, b, and c). This shift is more pronounced for R02 and R03. The NAO ridge extends from northwest Africa to northwest Europe, specifically exhibiting an eastward shift at lower latitudes. For cold spells over western regions of Europe (R04, R05, and R06), the gph anomaly at 500 hPa shows a positive anomaly located north of a negative anomaly (supplementary Fig. S3d, e, and f). This anomaly resembles a high-over-low North Atlantic blocking situation, which is similar to the findings of other studies (e.g., Trigo et al., 2004; Buehler et al., 2011; Pfahl, 2014). For R04 and R05, at high latitudes, the locations of the NAO ridge and EMed trough tend to shift toward the west of their respective climatological centers (Fig. 6d and e). During cold spells over eastern Europe (R08), the positions of the EMed trough and the TP ridge at high latitudes shift toward the east of their respective climatological locations (Fig. 6h). The cold spell over Northern Siberia (R09) primarily aligns with a deep trough downstream over northwest Asia, as depicted in the supplementary Fig. S3i and Fig. 6i. Despite the high latitude of this region, it is sufficiently far from the pole to become affected by cold advection from higher latitudes during cold spells

230 (Tuel and Martius, 2024). For R09 (Fig. 6i), ridges form over the western flank of the Ural Mountains, and the location of
the EAsia trough at high latitudes shifts toward the west of its climatological location. These findings indicate that a distinct
shifting pattern of ridges and troughs on a hemispheric scale occurs during cold spells over each region. These shifts can
also be observed at 500 hPa (not shown) and 850 hPa (not shown), indicating that during cold spells the entire troposphere
is shifting in a rather barotropic manner. These shifting patterns suggest an interaction between atmospheric dynamics and
235 surface temperatures.

Fig. 7 shows the retrograding patterns of ridges and troughs before and during cold spells for the composites of regions in
North America and Europe. This choice is based on the similarity of the upstream ridge and downstream trough in the selected
regions for each composite. In North America, cold spell regions are predominantly located in the vicinity of the WNA ridge.
In North America, before cold spells occur, the ridge encompasses a broad area, including the cold-spell regions (Fig. 7a).
240 At the start of cold spells, the upstream ridge becomes more localized in the western side of the cold-spell regions (Fig. 7e).
This pattern provides northerly wind flow from the Arctic toward the cold-spell area (supplementary Fig. S4). In Europe,
most of the cold spell regions are primarily located in the vicinity of the EMed trough. Similar to North America, before cold
spells occur, the ridge encompasses a wide area (Fig. 7b), indicating that cold or warm anomalies are not concentrated to
specific locations. However, with the onset of cold spells (Fig. 7f), ridges and troughs become more concentrated, with a ridge
245 positioned upstream and a trough located within the cold-spell regions (Fig. 7f). Hence, associated with cold spells, roughly
the same type of regional-scale circulation pattern is found in both North America and Europe. These findings highlight the
critical role of the nearest upstream ridge and downstream trough in the formation of cold spells.

4.2 Trough and ridge amplitude and speed

For cold spells, the largest ridge and trough amplitude and speed differences relative to climatology occur for the local upstream
ridge and downstream trough (supplementary Figs. S5 and S6). Generally, the amplitude and speed differences of ridges and
250 troughs decay as their distance increases to the location of the cold spell. A composite over all cold spells and regions of the
upstream ridges and downstream troughs provides a general overview of the local wave behavior during cold spells (Fig. 8).
The upstream ridge significantly amplifies (Fig. 8a and supplementary Fig. S5), facilitating the transfer of Arctic cold air to
the cold spell region. During cold spells, the northward winds entering the Arctic are increasing west of the upstream ridge,
255 while the southward winds exiting the Arctic are increasing east of the upstream ridge (supplementary Fig. S4). The upstream
ridge also significantly slows down (Fig. 8b and supplementary Fig. S6), implying the persistence of the warm northward wind
to the Arctic and the cold southward wind to the cold spell location. The persistent warm northward wind to the Arctic might
be the reason behind the correlation between cold events and a warmer Arctic found in the other studies (e.g., Cohen et al.,
2018); yet more research is needed to discover the impact of the amplified ridge on Arctic temperatures. At low latitudes, the
260 upstream ridge moves faster, which mainly results from cold spells in Northern America and northern Siberia (supplementary
Fig. S6). The underlying reasons for this pattern can be explored in future research.

The downstream trough amplifies (deepens) at latitudes north of the cold spell location, while it weakens at southerly
latitudes (Fig. 8a and supplementary Fig. S5). For the cold-spell regions in the southern as compared to the northern mid-

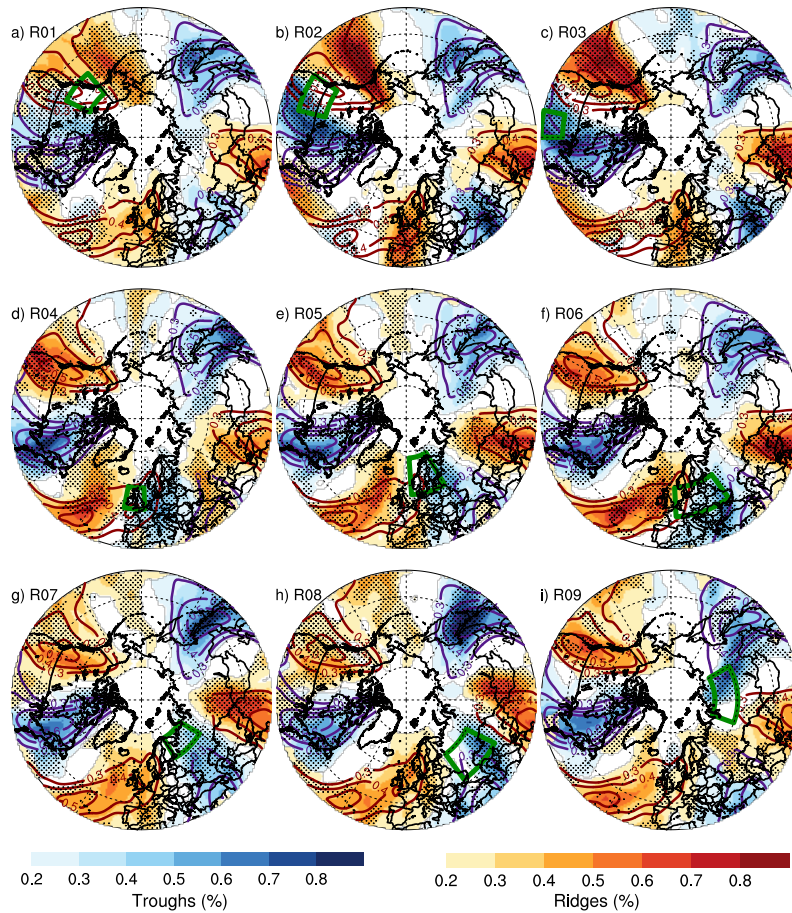


Figure 6. The probability of ridges and troughs at 300 hPa as a function of latitude and longitude. The shading (based on all-time steps of the cold spells) and contours (climatology) indicate probabilities of ridges and troughs in percentage. Solid brown (purple) contours represent the climatological probability of ridges (troughs). Red (blue) shading represents the probability of ridges (troughs) during cold spells over each region. As the probability of ridges and troughs are computed individually, shading and contours of the two fields are shown for values greater than 0.2% to minimize overlap in the plot. Dots indicate regions where cold spells and climatology are significantly different at the 95% confidence level. The green box in each plot shows the location of cold spells. A Gaussian function, using a weighted running average over eight degrees in latitude and over an equal distance in longitude, is used to smooth the probabilities. a) Northwest Canada; b) Northwest US and Northwest Canada; c) Central US; d) British Isles; e) Scandinavia; f) Europe; g) Northwest Russia; h) Southwest Russia; and i) Northern Siberia.

latitudes, the increase in trough amplitude develops more to the south (supplementary Fig. S5). Amplification of this trough
 265 contributes to enhancing the southward cold wind from the Arctic into the cold spell region (supplementary Fig. S4). Similar
 to the upstream ridge, the downstream trough significantly slows down (Fig. 8b and supplementary Fig. S6), hereby increasing
 the persistency of the cold southward winds into the cold spell region.

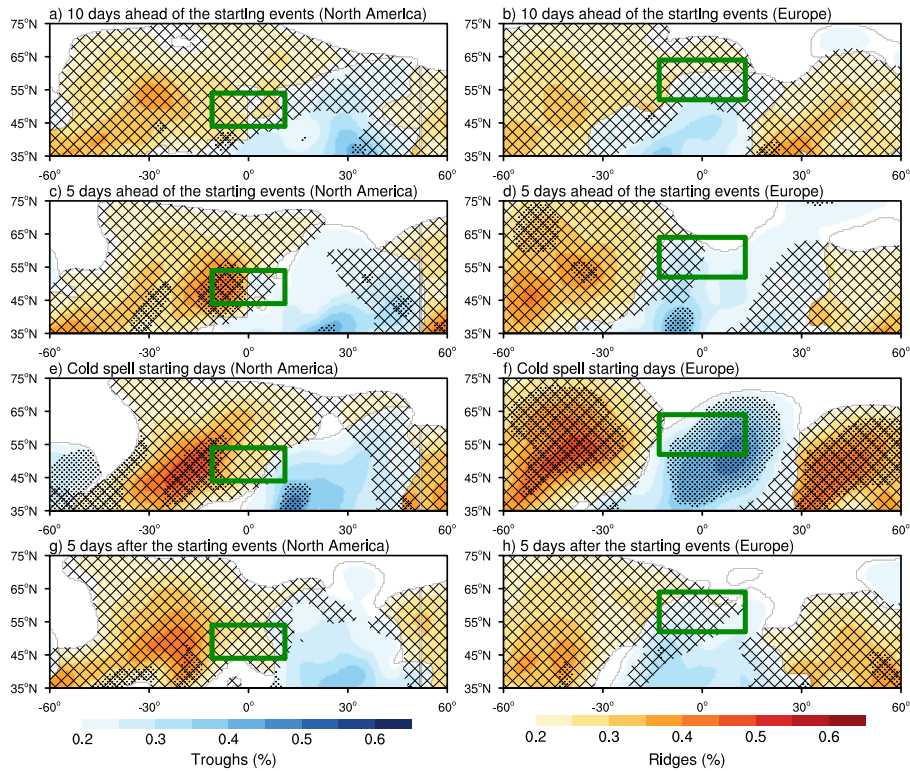


Figure 7. The probability of ridges and troughs at 300 hPa as a function of location and time relative to the start of the cold-spell events. Shown with shading are probabilities (in percentage) of ridges (red) and troughs (blue) for different time lags averaged over all cold spells in study regions relative to the center of the cold-spell region. The panels on the left correspond to the composite of cold spells in North America (R01, R02, and R03), and the panels on the right correspond to the composite of cold spells in Europe (R04, R05, R06, R07, and R08). To show both ridges and troughs in a single plot, priority is given to the stronger feature. The hatched areas represent grid points where ridges are occurring, and dots indicate regions where ridges and troughs during cold spells are significantly different from the climatology at the 95% confidence level. The green box shows the average latitude and longitude span of the cold spell regions. A Gaussian function, using a weighted running average over eight degrees in latitude and over an equal distance in longitude, is used to smooth the probabilities. a), b) 10 days ahead of the starting cold spells; c), d) 5 days ahead of the starting cold spells; e), f) cold spell starting days; and g), h) 5 days after the start of cold spells.

4.3 Lead-lag analysis

In the previous section, we have shown that cold spells are associated with higher-amplitude and slowing ridges and troughs in the vicinity of the cold spells. The question that arises is whether cold spells cause the amplification and slowdown of

waves or vice versa. Previous research has shown that upper-atmospheric changes—such as amplified planetary waves (Screen and Simmonds, 2014), blocking (Kautz et al., 2022), and the stratospheric polar vortex (Tomassini et al., 2012)—can lead to surface extreme events. Here, a day-lag analysis of composites over cold spells is provided to shed light on the causal relationship between upper-level waves and surface temperature (Fig. 9 and supplementary Fig. S7). Since the largest wave amplitude and speed differences relative to climatology belong to the upstream ridge and the downstream trough, their time series are plotted here.

The minimum speeds of the upstream ridge and downstream trough occur at a negative time lag several days prior to the onset of the cold spell, which is taken as 30 percent of grid points in the region experiencing a cold spell and indicated by the lag zero. The wave speed anomaly shows almost a similar pattern for both 500 hPa and 300 hPa (Fig. 9a and b). This pattern suggests an almost barotropic wave signal through the troposphere during cold spell events. However, the wave-speed reduction occurs somewhat later at 850 hPa, indicating a downward propagation of the wave-speed signal.

Both amplified ridge and trough are also encountered prior to the cold spell, but the speed reduction precedes the amplitude increase (Fig. 9a and b). This could potentially be attributed to the breaking of waves due to exceeding the capacity of the wave for the wave activity flux (Nakamura and Huang, 2018). The preceding of the upstream ridge and the downstream trough amplification relative to the cold spell suggests that cold spells arise as a result of slow and amplified local upper-atmospheric waves. For Northern Siberia, the downstream trough amplifies at a positive time lag several days after the cold spells (supplementary Fig. S7i), indicating that the amplified trough cannot be responsible for the cold spell in this region. This is consistent with the formation of cold spells in Northern Siberia being due to diabatic processes (Röthlisberger and Papritz, 2023). Nonetheless, the upstream ridge amplifies and slows down prior to the onset of cold spells in this region (supplementary Fig. S7i).

The upstream ridge maximizes simultaneously at all three levels, while the downstream trough amplifies almost simultaneously at 500 and 300 hPa and one day later at 850 hPa. This indicates that during cold spells, the upstream ridge amplification is barotropic, whereas in the lower troposphere, the downstream trough amplification is somewhat baroclinic. It has been shown that during cold spells over China, the positive gph anomaly has a barotropic structure and the negative gph anomaly has a baroclinic structure in the lower troposphere (Zhang et al., 2021).

There is no indication that the upstream ridge anomaly is encountered prior to the downstream trough anomaly or vice versa. This suggests that the anomalies could emerge independently in both the ridge and the trough. For instance, in January 2014, poleward low potential vorticity from subtropical deep convection amplified the WNA ridge, while equatorward and downward high potential vorticity deepened the ENA trough (Davies, 2015). Li et al. (2017) also showed that only the breakdown of the stratospheric polar vortex cannot be a reliable predictor of cold spells in North America, and warm sea surface temperature in the eastern extratropical Pacific also needs to be considered.

The lead-lag behavior of other significant and non-significant changes in the upstream ridge and downstream trough (supplementary Figs. S5 and S6) is also analyzed. Their time lag either shows no significant changes or occurs after the upstream ridge and downstream trough's significantly positive amplitude and negative speed anomalies (not shown). However, there are some exceptions: During cold spells over North America (R01, R02, and R03), the lead-lag behavior of the upstream ridge anomaly

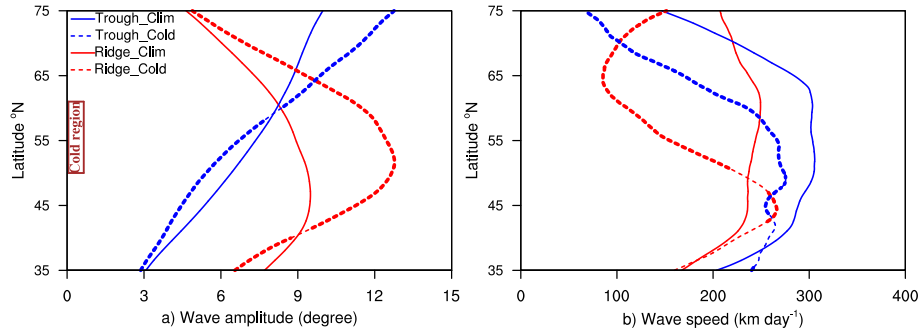


Figure 8. Climatology (solid lines) and composite of cold spells over all study regions (dashed lines) for a) wave amplitude (degree) and b) wave speed (km day^{-1}) at 300 hPa. Red and blue lines are showing the average of the upstream ridge and the downstream trough, respectively, in the vicinity of each cold spell region. Bold dashed lines indicate difference between cold spell and climatology at the 95% confidence level. The brown box on the left side shows the average latitude span of all regions. A four-degree running average is used to smooth the lines.

at the lower midlatitudes (35N-50N) indicates that this ridge slows down at these latitudes before it does at other latitudes (supplementary Figs. S8a). For both Eastern European regions (R07 and R08), latitudes with a reduction in the upstream ridge amplitude (35N-40N; supplementary Fig. S7g and h) show strong negative amplitude anomalies weeks before the onset of the cold spell (supplementary Figs. S8b). This indicates that the ridge at lower latitudes tends to move zonally, which may weaken
 310 the mixing of warm tropical air with colder midlatitude air long in advance of the cold spell formation.”

5 Conclusions

The present study provides a comprehensive perspective on the formation of cold spells in the midlatitudes, demonstrating that increased wave amplitude and decreased speed are fundamental drivers of cold-spell development. Across all midlatitude regions, locally amplified and slowing ridges and troughs near cold spells appear important for the formation of these.
 315 Through a daily lag analysis, a cause-and-effect relationship between upper-level waves and extreme cold surface temperatures is revealed, indicating that upper-level waves are preceding cold spells and hence important for the development of these.

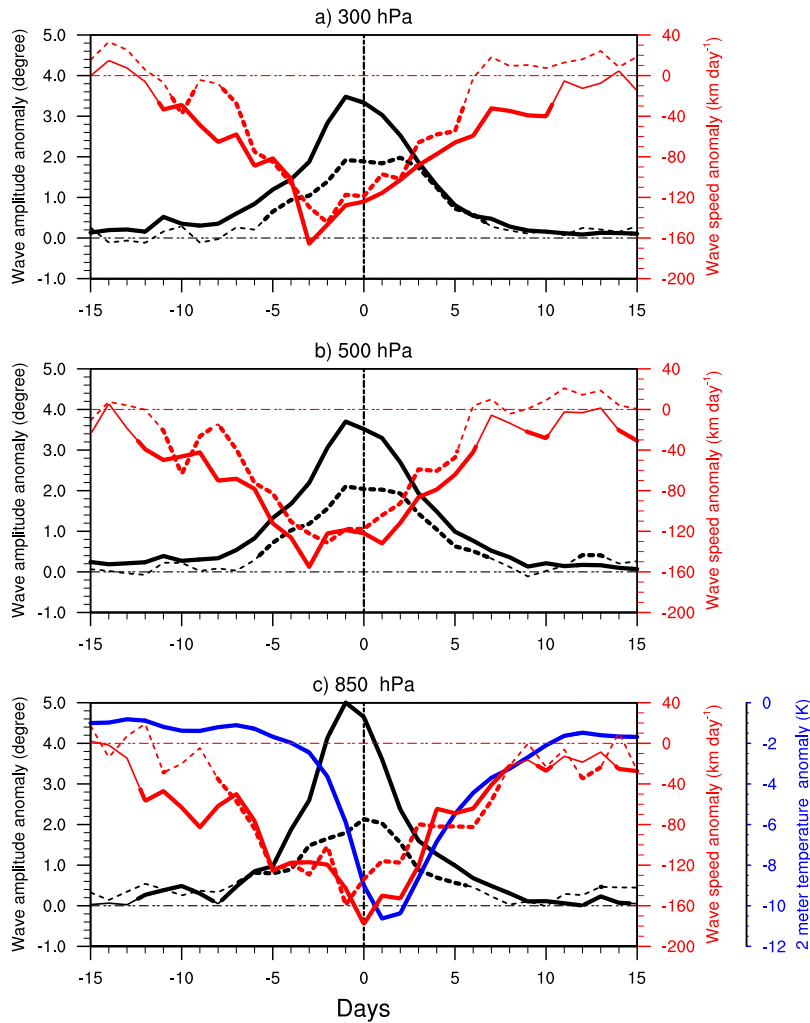


Figure 9. Composites of anomalies relative to climatology of 2-meter temperature, and ridge and trough speed and amplitude during all cold spells as a function of time lag for all study regions. The lag zero (zero on the x-axis) is the day that 30% of the land in each region experiences a cold spell. Blue line is the composite of temperature anomalies averaged over the whole study region. Black and red lines are the composites of ridge and trough amplitude and speed anomalies, respectively. The upstream ridge and the downstream trough in the vicinity of the cold-spell location are shown by solid and dashed lines, respectively. The latitude bands with a significant positive wave amplitude anomaly and a negative wave speed anomaly are averaged for each cold spell region (indicated by the vertical solid lines on the right side of the supplementary Figs. S5 and S6). Bolded solid and bolded dashed lines indicate the difference to climatology significant on a 95% level. a) 300 hPa, b) 500 hPa, and c) 850 hPa.

Our findings support previous research conducted in a specific region, indicating that slow (e.g., Fragkoulidis et al., 2018; Fragkoulidis and Wirth, 2020) and amplified (e.g., Jolly et al., 2021; Fragkoulidis and Wirth, 2020) Rossby waves contribute

to cold spells Europe. Moreover, we demonstrate the importance of ridge and trough development at each latitude within the
320 mid-latitudes, whereas others (e.g., Fragkoulidis and Wirth, 2020) discuss daily averages of speed and amplitude over an area
encompassing the cold-spell region. By focusing on latitude rather than averaging over a broader region, we determine that the
slow and amplified waves are mostly in the vicinity or north of the cold spell regions.

Additionally, we discuss the importance of wave location in cold spell development, as shifting the location of waves relative
to climatology can result in a cold air advection to the cold region. Our findings highlight the critical importance of wave
325 location, emphasizing the need to consider potential shifts in wave positions in a warming world. Such shifts could result in
more frequent extreme events in certain areas, impacting local climates.

It is important to note that this research primarily focused on the nearest ridge and trough relative to the location of cold
spells. However, an upper-level anomaly can propagate both downstream and upstream (e.g., Simmons and Hoskins, 1979).
Consequently, the anomaly that ultimately triggers the formation of cold spells may initially originate from a remote ridge or
330 trough, and further investigation is required to explore each region in greater depth.

Considering the increasing frequency of heat waves, floods, and droughts due to climate change (IPCC; Seneviratne et al.,
2021), there is a need to understand the dynamical drivers of them (Xu et al., 2024). In this context, we have developed two
tools designed to identify the amplitude and speed of atmospheric waves, which can be utilized in future research to unravel the
dynamical drivers of various extreme events. Furthermore, these metrics may help clarify the impact of AA on the waviness
335 and speed of Rossby waves. Our meridional wave amplitude metric provides a unique methodology compared to other metrics
(Geen et al., 2023), as it could evaluate changes in ridge and trough independently in a warming world.

Code availability. Codes are available at <https://doi.org/10.5281/zenodo.16761350>

Data availability. We obtained the ERA5 data from the ECMWF data server on pressure levels at <https://cds.climate.copernicus.eu/datasets/reanalysis-era5-pressure-levels?tab=download> and single levels at [https://cds.climate.copernicus.eu/datasets/reanalysis-era5-single-levels?](https://cds.climate.copernicus.eu/datasets/reanalysis-era5-single-levels?tab=download)
340 [tab=download](https://cds.climate.copernicus.eu/datasets/reanalysis-era5-single-levels?tab=download).

Author contributions. All authors conceived and designed the study. Material preparation, data collection and analysis were performed by MB. The first draft of the manuscript was written by MB with comments and revisions from RGG and JPS. All authors read and approved the final manuscript.

Competing interests. The authors declare no competing interests.

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