19 Cold-air outbreaks over East Asia associated with blocking highs

mechanisms and their interaction with the polar stratosphere

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19.1 Introduction

The East Asian winter monsoon is one of the factors that dominate regional climate over the Far East and western North Pacific. Cold air is advected into the mid-latitude Far East by the prevailing monsoonal northerlies between the Siberian high (SH) and the Aleutian low (AL). At the same time, poleward heat transport is stronger than in any other region in the Northern Hemisphere (Higuchi et al., 1991, Nakamura et al., 2002). By sustaining a sharp meridional temperature gradient around Japan and abundant supply of heat and moisture from the warm ocean to the dry, cold, continental air mass, the winter monsoon can also influence storm track activity over the North Pacific (Blackmon et al., 1977, Hoskins and Valdes, 1990, Nakamura, 1992, Nakamura et al., 2002).

Sub-seasonal intensification of the winter monsoon accompanies a cold surge or a cold-air outbreak, giving rise to an abrupt temperature drop, severe frost, freezing rain, and heavy snowfalls over the Far East (Boyle and Chen, 1987, Jeong et al., 2008). The influence of a cold surge often reaches as far south as southern China (Wu and Chan, 1997), and, rarely, it may even influence convective activity over the Maritime Continent (Ding and Krishnamurti, 1987). Many previous studies of cold-air outbreaks have focused on atmospheric circulation and lower-troposphere heat budgets (Boyle and Chen, 1987, Ding and Krishnamurti, 1987, Clark et al., 1999). Nevertheless, it has been pointed out that sub-seasonal monsoon variability is closely linked to upper-tropospheric wave-like circulation anomalies (Suda, 1957, Joung and Hitchman, 1982, Lau and Lau, 1984, Hsu and Wallace, 1985, Wu and Chan, 1997, Takaya and Nakamura, 2005a).

More recently, the influence of a blocking ridge that forms over the subpolar western North Pacific on sub-seasonal SH intensity has also been demonstrated (Takaya and Nakamura, 2005b).

The wave-like circulation anomalies over Eurasia may also be related to interannual SH variability via the Eurasian (EU) teleconnection pattern (Gong et al., 2001) defined by Wallace and Gutzler (1981) or the Scandinavian pattern (Bueh and Nakamura, 2007), identified by Barston and Livezey (1987). However, it has been pointed out that more zonally symmetric anomalies associated with the Arctic Oscillation or Northern Hemisphere annular mode (NAM; Thompson and Wallace, 1998, 2000, 2001) may influence interannual variability of the winter monsoon (Gong et al., 2001, Wu and Wang, 2002, Gong and Ho, 2004, Jhun and Lee, 2004, Jeong and Ho, 2005). The winter monsoon is also known to exhibit multi-decadal variations (Nakamura et al., 2002, Panagiotopoulos et al., 2005, Wang et al., 2009).

This chapter highlights recent progress in the study of the variability of the SH and East Asian winter monsoon. We focus on the interaction of the upper-level circulation anomalies with the surface air temperature (SAT) gradient, which is shown to be essential to the sub-seasonal amplification of the cold SH (Takaya and Nakamura, 2005a, 2005b). We also examine monsoon variability over interannual and decadal time scales, with particular attention given to the associated modulation of planetary waves. Because of their large upward group velocity, the planetary waves over the Far East can act as a vertical “connector” between the troposphere and the stratosphere (Nakamura et al., 2010). As a surface manifestation of tropospheric planetary waves, the East Asian winter monsoon varies seasonally and interannually with modulated activity and structure of the planetary waves (Chen et al., 2005, Wang et al., 2009, Takaya and Nakamura, 2013), which can have specific impacts on the Arctic stratosphere. As revealed by (among others) Koldstad et al. (2010), the SH variability and associated cold-air outbreaks into East Asia are linked to the variability of the stratospheric polar vortex.
19.2 Upstream influence on winter monsoon variability

Cold-air outbreaks into the mid-latitude Far East are known to follow sub-seasonal SH amplification that occurs in conjunction with a Rossby wave train propagating from the North Atlantic along the subpolar westerly jet (Suda, 1957, Joung and Hitchman, 1982, Lau and Lau, 1984). Interestingly, for the first application of his wave theory, Rossby et al. (1939) looked to sub-seasonal SH fluctuations in relation to variations of zonal-mean westerlies (index cycle). Even earlier, Ficker (1911) found a large-scale wave-like SAT pattern over Russia, which can be interpreted today as a surface manifestation of a Rossby wave train propagating at the tropopause level in the course of sub-seasonal SH amplification.

Using the composite analysis and application of a wave-activity flux for stationary Rossby waves formulated by Takaya and Nakamura (2001), Takaya and Nakamura (2005a, 2005b) confirmed the primary importance of the Rossby wave train and an associated upper-level blocking ridge over the Ural Mountains for extreme SH amplification and subsequent cold-air outbreaks into East Asia (Fig. 19.1). They refer to this process as the Atlantic-origin type or wave-train type of the SH amplification. This wave train resembles the Eurasian (EU) teleconnection pattern. Using a potential-vorticity (PV) inversion (Hoskins et al., 1985), Takaya and Nakamura (2005a) elucidated how an equivalent barotropic anticyclonic anomaly associated with the Ural blocking can strengthen the cold surface SH (Fig. 19.1). The blocking develops to the west of the near-surface cold air mass, which accumulates climatologically over northeastern Siberia. To the southwest of the cold air mass a strong temperature gradient is located over Mongolia. Acting on this SAT gradient, which is equivalent to the equatorward PV gradient from PV thinking, anomalous northeasterlies induced at the surface by the blocking-associated tropopause-level PV anomalies yield anomalous cold air advection. The cold air mass thus pulled southward out of its core region contributes to the amplification of the surface SH. Once developed by the upper-level influence, the surface cold anomaly behaves as a surface thermal Rossby wave (Gill, 1982), acting to move itself eastward along the surface baroclinic zone collocated with a sharp meridional SAT gradient. As a surface thermal Rossby wave, the anomalous cold air thus generated forms an anticyclonic anomaly (Hoskins et al., 1985), yielding anomalous cold advection to its east to

Fig. 19.1. Intra-seasonal intensification of the Siberian High (SH) composited for two days before the peak times of the 20 strongest sub-seasonal events observed around [47°N, 90°E] over 40 recent winters. (a) 250 hPa height anomaly (±50, ±150, ±250, ... m; dashed for negative) and associated wave-activity flux (arrows with scaling (units: m²s⁻²) at the lower right). (b) 1000 hPa height anomaly (±20, ±60, ±100, ... m; dashed for negative). In (a) and (b), SAT anomalies are superimposed (contoured for ±2, ±6, ±10, ...°C; stippled heavily and lightly for cold and warm anomalies, respectively). (c) Anomalous 1000 hPa wind (arrows with scaling (units: ms⁻¹) at the lower left) induced solely by composited 300 hPa PV anomalies, superimposed on total 1000 hPa temperature (contoured for every 8 °C and bold lines for 0 °C). (d) Anomalous tendency in 1000 hPa temperature (contoured for every 2 °C day⁻¹; dashed for warm advection) induced solely by the anomalous winds acting on the temperature field in (c). Observed warm and cold 1000 hPa temperature anomalies (stronger than 2 °C) are stippled lightly and heavily, respectively. The height anomalies and their tendencies are all rescaled with f(45°N)/f(lat.), and the compositing was based on eight-day low-pass-filtered fields. After Takaya and Nakamura (2005a), Nakamura et al. (2010), and based on the NCEP/NCAR reanalysis data set.
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shift the cold air itself eastward. Behind this eastward-moving cold anomaly, the upper-level blocking anticyclone continues to induce the anomalous cold advection, acting to retard the eastward migration of the cold anomaly and thereby retain its amplification. The anticyclonic anomaly thus built up has a baroclinic structure whose phase lines tilt northwestward with height, and it thereby transports heat poleward with a distinct upward component of Rossby wave-activity flux. More detailed discussions on the mechanisms through which the upper-level blocking anomaly intensifies the surface cold SH are given by Takaya and Nakamura (2005a) and by Nakamura et al. (2010).

Similar to this intra-seasonal connection, there is also a close connection between variations in the Ural blocking and SH at the interannual scale. In fact, Wang et al. (2010) found that the winter-mean Ural-blocking index exhibits a high positive correlation (+0.69) with the winter-mean SH intensity, which is significant at the 99% confidence level for the 44 winters since 1957. The winter-mean SH intensity, defined as sea-level pressure (SLP) averaged within [40–65°N, 80–120°E] as in Panagiotopoulous et al. (2005), is correlated with the Ural blocking more closely than with any other single teleconnection pattern investigated in previous studies. The Ural blocking typically accompanies an anticyclonic anomaly near the Kara Sea, and a pair of cyclonic anomalies, one over northern Europe and the other over East Asia. These wave-like anomalies are equivalent barotropic, and extend from the surface into the lower stratosphere. The blocking is associated with below normal and above normal lower-tropospheric temperatures over East Asia and northern Siberia, respectively, at the 99% confidence level, which result from the anomalous thermal advection around the blocking high.

19.3 Monsoon variability associated with the Western Pacific teleconnection pattern

In addition to the Rossby wave teleconnection from the Atlantic discussed above, intensification of the surface SH also follows a blocking formation over the subpolar western North Pacific (Takaya and Nakamura, 2005b), and the particular flow configuration resembles the positive phase of the western Pacific (WP) pattern (Pavan et al., 2000, Rivière, 2010). As one of the major atmospheric teleconnection patterns over the wintertime Northern Hemisphere (Wallace and Gutzler, 1981), the WP pattern is characterized by north–south dipolar height anomalies over the western North Pacific. In this particular type of SH amplification, which may be called the Pacific-origin type (Takaya and Nakamura, 2005b), the SH centre is shifted to the northeast of its climatological mean position. Unlike the Atlantic-origin type, there is no indication of wave train propagation to the amplifying blocking ridge from upstream (Fig. 19.2). Rather, the ridge develops as an anticyclonic anomaly and slowly retrogresses from the Aleutian region into eastern Siberia (Branstator, 1987, Kushnir, 1987). This process may be viewed as local inward (cyclonic) breaking of the polar vortex associated with the blocking formation (Swanson, 2000, 2001), because low-PV anomaly penetrates into the polar vortex and then becomes isolated within it, while high-PV air associated with the climatological trough over the Far East becomes elongated to the south of the blocking ridge. The development and maintenance of the blocking ridge occur under feedback forcing from synoptic-scale eddies migrating along the poleward-deflected Pacific storm track through their anomalous transport of vorticity (Takaya and Nakamura, 2005b). In contrast, the anomalous eddy heat transport acts to keep the blocking anomaly equivalent barotropic, especially during its amplification stage, even near the surface where the anticyclonic anomaly is accompanied by a warm anomaly. A distinct exception is observed to the northeast of the Tibetan Plateau, where a cold anticyclonic anomaly forms near the surface as the upper-level blocking anomaly is anchored to the north of the Sea of Okhotsk. It was revealed through a PV inversion diagnosis (Takaya and Nakamura, 2005b) that the upper-level blocking anomaly induces anomalous northeasterlies that advect an extremely cold air mass accumulated climatologically over eastern Siberia towards the northeast of the Plateau. After the peak time, the near-surface cold anticyclonic anomaly extends south into the mid-latitude Far East, where a cyclonic anomaly associated with the WP pattern still remains in the upper troposphere. Such baroclinic structure of the circulation anomalies over eastern China and Japan associated with the anomalous cold surface air is common to both the Atlantic-origin and Pacific-origin types of extreme SH amplification.

As one of the major factors that influence the East Asian winter monsoon (Takaya and Nakamura, 2005a, 2005b), the WP pattern significantly modifies the seasonal evolution of the tropospheric planetary waves (Takaya and Nakamura, 2013). In the upper troposphere the climatological seasonal progression from November to January over the extratropical Northern Hemisphere is characterized by an overall decline in geopotential height associated with tropospheric cooling (not shown). However, the height decline is not uniform over the extratropics. Over the Pacific, the decline is particularly enhanced in a zonally elongated mid-latitude domain from the Far East into the eastern North Pacific. In contrast, over the subpolar region between central Siberia and Alaska, the corresponding height decline is much reduced, and the climatological December–January tendency is even slightly positive around the Bering Strait (Figs. 19.2–19.3). These height
Fig. 19.2. Composite time evolution for the 20 strongest blocking events at the 250 hPa level observed around a target grid point [67°N, 140°E] over 40 recent winter seasons that induced the intra-seasonal SH intensification. The compositing was performed relative to the peak time (day 0) for each of the events. (left) 8-day low-pass-filtered 1000 hPa height anomalies normalized by a factor of $\sin(45^\circ \text{N})/\sin(\text{lat.})$, contoured every 40 m from ±20 m (dashed for negative values). The anomalies significant at the 95% confidence level are stippled. (middle) 8-day low-pass-filtered temperature anomalies at the lowest model level ($\sigma = 0.995$) (contoured every 4 °C from ±2 °C; solid and dashed lines for cold and warm anomalies, respectively). Surface elevation over 1500 m is shaded. (right) 8-day low-pass-filtered 250 hPa height anomalies rescaled with $\sin(45^\circ \text{N})/\sin(\text{lat.})$, contoured every 100 m from ±50 m (dashed for negative values). The horizontal component of wave-activity flux is superimposed with arrows whose scaling (units: m²s⁻²) is given near the lower right panel. After Takaya and Nakamura (2005b) and based on the NCEP/NCAR reanalysis data set.

Fig. 19.3. Tendency in 250 hPa height (m) from November to January with colouring convention given below (c). Composed for 10 Januaries of (a) enhanced, and (b) weakened monsoon activity, and (c) climatological tendency. After Takaya and Nakamura (2013) and based on the NCEP/NCAR reanalysis data set.
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19.4 Long-term modulation of the Siberian high and East Asian winter monsoon

The climate system over the Far East and North Pacific underwent a significant shift in the mid 1970s in association with the Pacific Decadal Oscillation (Nitta and Yamada, 1989, Trenberth and Hurrell, 1994, Schneider and Cournuelle, 2005). A comparison between two periods, one before (1957–1976) and one after (1977–2000) this climatic shift, suggests possible modulations in seasonal impacts of the Ural blocking on the East Asian winter climate. During the earlier period, 850 hPa temperature anomalies over East Asia associated with the blocking were confined mainly to the north of 40°N (Fig. 19.5a), while the corresponding anomalies in the later period spread as far southeastward as about 25°N (Fig. 19.5b). Furthermore, the correlation coefficient between the winter-mean indices of the Ural blocking and SH intensity increased from +0.60 in the earlier period to +0.74 in the later period, exceeding the 99% confidence level in both periods. Correspondingly, the fraction of the interannual SH variance explained by the Ural blocking variability had increased from only 36% to 55%.

The aforementioned multi-decadal modulations were accompanied by enhanced downstream influence of the Ural blocking with a stronger cyclonic anomaly over East Asia in the mid-tropospheric wave-like anomalies associated with the blocking (Fig. 19.5c–d). This enhancement of

![Fig. 19.4. Polar stereographic maps (poleward of 30°N) of (a) 250 hPa and (b) 30 hPa height anomalies (grey lines for every 50 m; dashed for negative; zero lines omitted) composited for the peak times of the 18 strongest positive events of the Western Pacific (WP) pattern. Based on the JRA-25 reanalysis. (c and d) As (b), but for (c) +5 and (d) +20 days, respectively, relative to the peak time. A black thick line in (a) indicates 250 hPa height of 10,000 m. See Nishii et al. (2010) for the corresponding detailed plots.]
the downstream extension may be explained partly from the viewpoint of stratosphere–troposphere dynamical coupling (Wang et al., 2010). The Ural blocking accompanies a lower-stratospheric anticyclonic anomaly over the Arctic region and cyclonic anomalies in mid latitudes, and this vertical coupling had weakened into the later period with a reduction in statistical significance (Fig. 19.5e–f). This feature is indicative of a compensatory tendency.
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between the upward influence of the Ural blocking into the stratosphere and its downstream influence into East Asia, which is suggested by the 3D propagation of stationary Rossby wave activity (Wang et al., 2010). This compensatory tendency may be related to multi-decadal modulations of the stratospheric polar night jet (PNJ) and associated polar vortex. During the earlier period, the polar vortex was in its weak regime (Christiansen, 2003), which favoured the upward propagation of planetary waves, perhaps including the wave propagation from the Ural blocking ridge. In the later period, the PNJ and stratospheric polar vortex strengthened (Christiansen, 2003), and this suppressed the upward propagation of planetary waves. Consequently, the stationary Rossby wave activity emanating from the Ural blocking may become more confined to the troposphere, leading to a stronger downstream influence on SH intensity.

The intensity of the East Asian winter monsoon is known to show multi-decadal variations (Shi, 1996, Nakamura et al., 2002, Panagiotopoulos et al., 2005). An index defined by Wang and Chen (2014b) identifies three distinctive periods of monsoon intensity: two strong epochs, one from the mid 1970s to the mid 1980s and the other since the early 2000s, separated by a weak epoch. The multi-decadal variations of the winter monsoon were associated with modulations of the planetary waves and the intensity of the Pacific jet stream (Nakamura et al., 2002). Compared with the strong epochs, the weak epoch was characterized by a weaker planetary-wave trough over the Far East and the Pacific jet, enhanced eddy activity along the Pacific storm track in midwinter (Nakamura et al., 2002), and was accompanied by higher winter-mean SAT over East Asia (Wang and Chen, 2014a), a reduced likelihood of cold-air outbreaks into China (Wang et al., 2009), enhanced precipitation over South China, and reduced precipitation over the mountainous regions of northern Taiwan (Hung and Kao, 2010, Wang and Feng, 2011).

In addition to these distinct climatic changes associated with multi-decadal variations of the winter monsoon, there are subtle but significant differences between the two strong epochs. During the earlier epoch (1976–1987), a significant negative SAT anomaly as strong as −1 °C was observed, mainly along the East Asian coast (Fig. 19.6a), but also in northwestern Europe, southeastern North America, and the central North Pacific, while no significant warming was observed in the extratropical Northern Hemisphere. In contrast, during the recent strong epoch (2004–2012) a significant negative SAT anomaly, which exceeds −3 °C, is extending zonally, but confined to the inland area of northern East Asia (Fig. 19.6d), while significant positive SAT anomalies are observed over the Arctic and the Tibetan Plateau. Circulation anomalies observed during the earlier epoch (Fig. 19.6b–c) resembled those associated with NAM, but those from the recent epoch do not. While exerting some influence (Jhun and Lee, 2004, Jeong and Ho, 2005, Wang et al., 2009), NAM may not necessarily be the controlling factor of the decadal variability in the winter monsoon. However, a common feature between the two epochs is a significant barotropic anticyclonic anomaly around the Ural Mountains (Fig. 19.6b, c, e, and f), suggestive of the crucial contribution of the Ural blocking to the multi-decadal variations of the winter monsoon. In fact, the Ural blocking frequency decreased in around 1985, and then increased again in around 2002 (Barriopedro et al., 2006, Wang and Chen, 2014a).

19.5 Summary and discussion

This chapter has considered recent progress in the study of the variability of the SH and East Asian winter monsoon. The interaction between upper-level circulation anomalies and a blocking anticyclone with a SAT gradient, which induces anomalous cold advection to intensify the cold SH (Takaya and Nakamura, 2005a, 2005b), is the main focus of the chapter. This is the mechanism through which the cold surface anomaly is induced by a warm, equivalent barotropic anomaly associated with a blocking anticyclone developing either over the Ural Mountains in association with a Rossby wave train from the Atlantic, or over the subpolar western North Pacific associated with the positive phase of the WP pattern.

Consequently, we also considered monsoon variability at interannual and decadal time scales from the viewpoint of modulations of the planetary waves. We concentrated on the monsoon variability associated with the WP pattern, whose positive phase corresponds to a blocking flow configuration over the Far East and western North Pacific. The seasonal tendency towards the positive WP pattern into midwinter leads to intensification of the winter monsoon associated with the enhanced seasonal development of planetary waves over the mid-latitude Pacific sector. Conversely, the blocking flow configuration leads to breaking of the planetary wave trough over the Far East and its weakening at subpolar latitudes, resulting in the reduction in upward propagation of the planetary waves and thereby the strengthening/cooling of the stratospheric polar vortex. If superimposed on the cooling trend in the stratosphere associated with global warming, this dynamically forced stratospheric cooling may enhance ozone depletion in the Arctic stratosphere. However, as noted earlier, the influence of a blocking high on the stratosphere is sensitive to its geographical location relative to the geographical phase of the climatological planetary waves (Woollings et al., 2010, Nishii et al., 2011). Thus, a blocking high that develops in association with a wave train from the Atlantic for SH intensification acts to weaken the stratospheric...
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Fig. 19.6. Composited winter-mean anomalies in (a) surface air temperature (SAT; every 1 °C), (b) sea-level pressure (SLP; every 1 hPa), and (c) 500 hPa geopotential height (every 10 gpm) for the earlier epoch (1976–1987) of the strong East Asian winter monsoon based on the NCEP/NCAR reanalysis data set. The negative anomalies are defined as the deviations from the corresponding composites for the epoch (1988–2003) of the weak monsoon. Negative anomalies are indicated with dashed lines. Dark and light shading indicates the 99% and 95% confidence levels, respectively. (d)–(f) are as (a)–(c), but for the corresponding later epoch (2004–2012) of the strong monsoon. After Wang and Chen (2014a).
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polar vortex (Nishii et al., 2011). In fact, a significant anticyclonic anomaly has been identified around the Ural Mountains as a robust precursory signal in the troposphere for the NAM-related weakening of the stratospheric polar vortex (Takaya and Nakamura, 2008, Koldstad et al., 2010).

It should be emphasized that the linkage between the winter monsoon variability and stratospheric variability is not necessarily a one-way process from the former to the latter. In fact, the weakening of the stratospheric polar vortex tends to be translated into the troposphere, acting to set up conditions favourable for cold-air outbreaks into mid latitudes (Thompson and Wallace, 2001, Koldstad et al., 2010). The influence of the Ural blocking on a cold-air outbreak into East Asia may depend on the strength of the stratospheric PNJ (Wang et al., 2010). Furthermore, Kodera et al. (2013) argued that, in same cases, a blocking high over the western North Pacific can develop in association with a planetary wave packet that has been reflected back from the stratosphere. Further study is required to deepen our understanding of the two-way nature of the monsoon-stratosphere linkage and to assess whether this linkage can be used to improve predictions of monsoon variability.

One may wonder what forces a wave train over northern Eurasia or the positive WP pattern that induces the winter monsoon variability. As it tends to be observed during the cold phase (La Niña) of El Niño/Southern Oscillation (ENSO; Horel and Wallace, 1981), the positive WP pattern provides a linkage between the Eastern Asian winter monsoon and ENSO. Likewise, decadal-scale anomalous coolness in sea surface temperature (SST) in the tropical Pacific may set a condition favourable for the formation of the positive WP pattern. In addition to these remote influences from the tropics, SST anomalies in the mid-latitude North Pacific may also force anomalies similar to the WP pattern that influences the polar stratosphere, as suggested by a recent numerical study (Hurwitz et al., 2012). Although not discussed in detail here, Pacific SST anomalies either in the tropics or mid latitudes can also change the intensity of the Aleutian low (Horel and Wallace, 1981, Nitta and Yamada, 1989, Trenberth and Hurrell, 1994, Taguchi et al., 2012), through which the winter monsoon may be modulated (Nakamura et al., 2002). Recent studies have also suggested that rapidly declining ice cover in the Kara/Barents seas in late autumn may have resulted in the recent intensification of the winter monsoon and associated cold conditions prevailing over Eurasia (Honda et al., 2009, Inoue et al., 2012). A stationary Rossby wave train that acts to intensify the SH may be forced directly by enhanced sensible heat release from the ocean due to reduced ice cover, and indirectly through feedback forcing from synoptic-scale cyclones and anticyclones migrating along a deflected storm track. Further study is necessary to clarify the mechanisms through which these boundary-forcing mechanisms can generate teleconnection to the winter monsoon, and this improved understanding could be used to increase the predictability of monsoon variability. In addition, further study is also required to establish how global warming will alter the winter monsoon. Global warming itself should ultimately weaken the winter monsoon, but in the course of its weakening, the monsoon may temporarily intensify, for example, under the influence of the decline in sea ice cover in the Arctic.

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