Stratospheric variability contributed to and sustained the recent hiatus in Eurasian winter warming

Chaim I. Garfinkel1,2, Seok-Woo Son2, Kanghyun Song2,3, Valentina Aquila3, and Luke D. Oman4

1The Fredy and Nadine Herrmann Institute of Earth Sciences, Hebrew University of Jerusalem, Jerusalem, Israel, 2School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea, 3Department of Earth and Planetary Science, The Johns Hopkins University, Baltimore, Maryland, USA, 4NASA Goddard Space Flight Center, Greenbelt, Maryland, USA

Abstract The recent hiatus in global-mean surface temperature warming was characterized by a Eurasian winter cooling trend, and the cause(s) for this cooling is unclear. Here we show that the observed hiatus in Eurasian warming was associated with a recent trend toward weakened stratospheric polar vortices. Specifically, by calculating the change in Eurasian surface air temperature associated with a given vortex weakening, we demonstrate that the recent trend toward weakened polar vortices reduced the anticipated Eurasian warming due to increasing greenhouse gas concentrations. Those model integrations whose stratospheric vortex evolution most closely matches that in reanalysis data also simulate a hiatus. While it is unclear whether the recent weakening of the midwinter stratospheric polar vortex was forced, a properly configured model can simulate substantial deviations of the polar vortex on decadal timescales and hence such hiatus events, implying that similar hiatus events may recur even as greenhouse gas concentrations rise.

1. Introduction

The decade of the 2000s was the warmest decade in the satellite era and was significantly warmer than the 1990s or any other prior decade [Stocker et al., 2013]. However, there was a slowing in the rise of global-mean surface temperature in the 2000s [Easterling and Wehner, 2009], often referred to as a hiatus [Trenberth and Fasullo, 2013]. This recent hiatus in global-mean surface temperature warming included a pronounced Eurasian winter cooling trend (Figure 1a) [Cohen et al., 2012, 2014; Li et al., 2015; Kug et al., 2015]. While changing sea surface temperatures (SSTs) can largely account for the hiatus in other seasons and in other regions [Kosaka and Xie, 2013; Trenberth et al., 2014; England et al., 2014], they cannot explain the midwinter cooling over Eurasia [Kosaka and Xie, 2013; Thorne et al., 2015], and the cause(s) for this cooling is unclear [Li et al., 2015; Thorne et al., 2015]. Here we link the observed Eurasian warming in a recent trend toward weakened stratospheric polar vortices.

The midwinter stratospheric polar vortex has weakened significantly over the past several decades. Polar lower stratospheric geopotential height has risen in three different modern reanalyses products since 1980 and especially since 1990 (Figure 1b), and the linear trend over the full 35 years is significant at the 5% level. The linear trend in lower stratospheric polar cap temperatures and lower stratospheric wave flux in midwinter over the satellite era also indicates warming and weakening of the vortex [Cohen et al., 2009; Garfinkel et al., 2015] and since 1990 the polar lower stratospheric temperature trend is significant at the 5% level (e.g., 1.7 ± 1.5 K per decade at 150 hPa in ERA-Interim [Dee et al., 2011] and 2.1 ± 1.5 K per decade at 150 hPa in Modern-Era Retrospective-Analysis for Research and Applications (MERRA) [Rienecker et al., 2011]). This is likely related to decadal variability in stratospheric sudden warming frequency for most definitions of sudden warmings considered by Butler et al. (2015). It is well established that on subseasonal timescales, a warmer polar stratosphere and stratospheric sudden warmings are associated with cold temperatures over Eurasia [Thompson et al., 2002; Kolstad et al., 2010; Tomassini et al., 2012; Lehtonen and Karpechko, 2016], and this downward coupling effect is independent of any prior memory in the troposphere and hence enhances surface climate predictability [Charlton et al., 2004; Gerber et al., 2009; Sigmund et al., 2013; Tripathi et al., 2015; Scaife et al., 2016]. Nakamura et al. (2016) suggest that a similar connection may exist on interannual and decadal timescales. Can this effect, along with the observed vortex weakening trend, explain the recent lack of winter surface warming over Eurasia?
Figure 1. (a) Evolution of Eurasian surface temperature in MERRA (Modern-Era Retrospective-Analysis for Research and Applications) [Rienecker et al., 2011], JRA55 (Japanese 55 year Reanalysis) [Kobayashi et al., 2015], and ERA-I (ERA-Interim) Dee et al., 2011] reanalysis data in December to February (DJF). (b) Variability in lower stratospheric (150 hPa) polar cap (70°N and poleward area averaged) geopotential height in MERRA, JRA55, and ERA-I reanalysis data. (c) As in Figure 1a but regressing out the influence of lower stratospheric polar cap geopotential height (shown in Figure 1b). The trend since 1990 is indicated in gray. Eurasian surface temperature is defined in section 3. Variability in globally averaged 2 m temperature over land in MERRA and ERA-I reanalysis data is shown in supporting information Figure S1. Confidence intervals are based on a $p$ value of 5% (note that the trend in Figure 1a would be considered significant if one accepted a $p$ value of 10%.

2. Data

We answer this question by analyzing MERRA [Rienecker et al., 2011], JRA55 (Japanese 55-year Reanalysis) [Kobayashi et al., 2015], and ERA-Interim [Dee et al., 2011] reanalysis data and output from atmospheric chemistry-climate general circulation models (GCMs) and coupled ocean-atmosphere GCMs. The atmospheric chemistry-climate GCM output is from a 29 member ensemble using the Goddard Earth Observing System Chemistry-Climate Model, Version 2 (GEOSCCM) [Rienecker et al., 2008] of the period from January 1980 to December 2009, though some integrations extend for longer. GEOSCCM couples the Goddard Earth Observing System Model, Version 5 [Rienecker et al., 2008; Molod et al., 2012] atmospheric GCM to the comprehensive stratospheric chemistry module StratChem [Pawson et al., 2008]. The model has 72 vertical layers, with a model top at 0.01 hPa, and all simulations discussed here were performed at 2° latitude × 2.5° longitude horizontal resolution. Reanalysis data are interpolated to the same 2° latitude × 2.5° longitude degree grid used for the GEOSCCM simulations. These simulations have been performed for various purposes and differ in the forcings included and in the physical parameterizations, but they all include (at least) changing SSTs, sea ice, and greenhouse gas concentrations. We use time varying observed SSTs and sea ice up to November 2006 from the Met Office Hadley center observational database [Rayner et al., 2006] and from
the National Climatic Data Center [Reynolds et al., 2002] since then. Greenhouse gas concentrations are from observations up to 2005 and from the Representative Concentrations Pathway 4.5 after 2005 [Meinshausen et al., 2011]. Changing ozone depleting species concentrations track those observed for all but three integrations (for these, concentrations are fixed at 1960 values) though all results in this paper are similar if we remove these three.

In addition, we also analyze the relationship between stratospheric variability and Eurasian surface temperature on interannual timescales in CMIP5 model historical integrations with a model lid above 0.1 hPa. The models considered are CMCC-CESM, CMCC-CMS, GFDL-CM3, HadGEM2-CC, IPSL-CM5A-LR, IPSL-CM5A-MR, IPSL-CMSB-LR, MIROC-ESM-CHEM, MIROC-ESM, MPI-ESM-LR, MPI-ESM-MR, and MRI-CGCM3. We include all available ensemble members for each model, and in our presentation of the results, we do not distinguish between individual CMIP5 models. A total of 33 CMIP5 simulations are considered, though results are similar if we only select one ensemble member for each model. Note that these simulations end in 2004, and hence, we only consider the connection on interannual timescales between Eurasian surface temperature and the stratospheric state. It has already been shown that these integrations do not capture the hiatus [Meehl and Teng, 2014]. Note that low-top CMIP5 models fail to realistically simulate stratospheric variability [Charlton-Perez et al., 2013] and struggle to capture the downward coupling to the surface [Manzini et al., 2014]; we therefore do not consider them in this study.

3. Methods

Motivated by the recent evolution in Eurasian surface temperature and in the polar stratosphere evident in Figure 1, we consider trends from 1990 to 2009 in this paper. Trends and best fit lines are calculated with a linear least squares fit. Statistical significance of the trends and best fit lines are computed using a two-tailed Student’s \( t \) test. For trend estimates, the reduction in degrees of freedom due to autocorrelation of the residuals is taken into account with the formula \( N^{\frac{1}{1+r_1}} \), where \( N \) is the number of years and \( r_1 \) is the lag-1 autocorrelation [Santer et al., 2008, equation (6)]. For lagged correlations, the reduction in degrees of freedom is taken into account with the formula \( N^{\frac{1}{1+r_1+r_2}} \) as in Bretherton et al. [1999, equation (31)] where \( N \) is the number of years and \( r_2 \) and \( r_1 \) are the two indices whose correlation is computed. The net effect is that for the lagged correlation between Eurasian 2 m temperature and polar vortex variability, there are 87 degrees of freedom in the reanalysis record since 1979 and correlations higher than 0.211 are statistically significant at the 5% level by a Student’s \( t \) test. Use of equation (1) of Pyper and Peterman [1998] yields 149 degrees of freedom, and so lagged correlations above 0.161 are deemed statistically significant at the 5% level by a Student’s \( t \) test. Note that the skewness of Eurasian surface temperature as defined here is 0.04 for MERRA and 0.01 in ERA-I, and hence, the concerns raised in Garfinkel and Hamik [2016] regarding the validity of a Student’s \( t \) test are not relevant here.

Eurasian near surface temperatures are computed as an area-weighted average of the 2 m temperature at all land grid point from 40°N and poleward from 50°W to 180°E. Geopotential height at 150 hPa area averaged from 70°N and poleward is used to quantify the state of the polar lower stratosphere. Polar cap averaged geopotential height and an empirical orthogonal function-based northern annular mode index are indistinguishable [Baldwin and Thompson, 2009]. Results are similar, though generally weaker, if we use polar cap heights higher in the stratosphere than 150 hPa. Note that Baldwin et al. [2003] found that the optimum single level for forecasting the Arctic Oscillation on intraseasonal timescales is 150 hPa. Except for supporting information Figure S1, all analysis in this paper focuses on boreal winter from December through February (DJF).

When considering the relationship on interannual timescales between stratospheric and Eurasian surface variability, we first detrend the data by subtracting the linear trend over the period 1980 to 2009 for GEOSCCM and reanalysis data and over the period 1980 to 2004 for CMIP5.

4. Results

We start by comparing polar cap geopotential height at 150 hPa and Eurasian near surface temperatures poleward of 40°N in DJF for each data source in Figure 2. Figure 2a focuses on the relationship between these two metrics on decadal timescales in the 29 GEOSCCM integrations and in reanalysis data. The \( x \) axis presents the trend in polar cap geopotential height, and the \( y \) axis presents the trend in Eurasian temperatures at 2 m for that same data source. There is clearly a close relationship between these quantities (correlation of \(-0.81\))
Figure 2. Connection between polar cap stratospheric variability and Eurasian surface temperatures on (a) decadal and (b and c) interannual timescales in winter (December through February) in GEOSCCM and in the MERRA and ERA-I reanalyses. The JRA55 reanalysis is also included in Figure 2a. The abscissa indicates the variability in polar cap geopotential height at 150 hPa, while the ordinate indicates variability in Eurasian surface air temperatures from 40°N and poleward. Figure 2c is the same as Figure 2b but for the 33 CMIP5 integrations with a high top. Red diamonds in Figure 2a mark the three hiatus ensemble members used for Figure 3. Confidence intervals on best fit slopes are based on a p value of 5%.

such that reduced decadal-scale warming over Eurasia occurs in simulations with a decadal-scale weakening of the stratospheric polar vortex (and vice versa), and the trends in the reanalyses fall along the best fit curve. A null hypothesis of no connection between these two quantities can be rejected at the 1% level using a Student's t test. If we choose 100 hPa polar cap height instead of 150 hPa, the correlation is −0.76. Hence, we can conclude that stratospheric variability is associated with the hiatus.

Our next step is to quantify this association. We do this by quantifying the connection between these metrics on decadal and interannual timescales for all of our data sources. Figure 2b focuses on the relationship between these two metrics on interannual timescales. On these timescales as well there is a strong relationship between the two (correlation of −0.55) as expected from previous work [e.g., Thompson et al., 2002]. The slope of the best fit linear line is indicated on the panel as well, and the slopes are indistinguishable on both interannual and decadal timescales, and also for both GEOSCCM and reanalysis data.

A similar comparison can be made for those CMIP5 models which fully resolved the stratosphere (Figure 2c). In the CMIP5 models, as in GEOSCCM, there is a strong relationship between stratospheric variability and Eurasian temperatures on interannual timescales (correlation of −0.53). The slope of the best fit linear line is indicated on the panel as well, and the slope is very similar to that in reanalysis data and in GEOSCCM: a 100 m change in 150 hPa polar cap height is associated with approximately 1 K of surface warming over Eurasia in all data sources. The similarity of the slope on interannual timescales among GEOSCCM, the CMIP5 models, and reanalysis indicates that this relationship is robust and that we may be able to use this slope to quantify the portion of the hiatus in Eurasian surface warming linked to stratospheric variability.
Using these best fit lines, we can quantify the possible influence that the trend toward higher polar cap heights (i.e., a weaker vortex) had on surface air temperature. Specifically, we linearly regress out the contribution of polar stratospheric variability shown in Figure 1b from the evolution of Eurasian 2 m temperatures shown in Figure 1a using a slope of $-1.13 \text{ K (100 m)}^{-1}$ slope (this slope leads to zero correlation between polar cap height and the adjusted surface temperature). See Figure 1c for the adjusted surface temperature over Eurasia. First, it is worth noting that the temperature evolution in Figure 1c is smoother than that in Figure 1a—the standard deviation drops from 1.35 K to 0.93 K. More importantly, Eurasian temperatures warm over this period by $0.23 \pm 0.49 \text{(decade)}^{-1}$ in Figure 1c instead of a near-statistically significant cooling trend in Figure 1a. This rate of warming quantitatively matches the rate of global warming as calculated by Foster and Rahmstorf [2011], suggesting that the recent stratospheric polar vortex weakening is linked to the lack of the surface warming even as greenhouse gas concentrations increase.
Next, we consider the three GEOSCCM members in which Eurasian surface temperatures cool over this period (indicated with a red diamond in Figure 2a). Figure 3 shows a map view of the surface air temperature trends in MERRA, ERA-I, and in these three GEOSCCM integrations. Figures 3a–3c show that in the annual average, the hiatus is not evident in all data sets. However in DJF, we see cooling over Eurasia in both reanalysis and ensemble mean of the three GEOSCCM runs (Figures 3d–3f). Figures 3g–3i show trends in each of the three hiatus members separately. The precise pattern of Eurasian cooling differs among the three, but all show large regions of cooling. These three integrations all simulated a trend toward a weaker polar vortex (Figure 2a).

Globallandsurfacetemperaturesintheseensemblemembersalsocloselytrackobservedlandsurfacetem- (supporting information Figure S1)—i.e., these simulations capture both the cooling in Eurasia in winter and the lack of warming in the annual average.

Thus far, we have demonstrated tight coupling between stratospheric variability and surface air temperature variations over Eurasia. However, one might ask about the direction of causality: did the weakened stratospheric polar vortex lead to the Eurasian surface cooling or vice versa? We answer this question by considering two metrics of the connection between stratospheric variability and Eurasian surface air temperature variability on intraseasonal timescales.

The first metric is the lagged correlation between these two diagnostics in DJF (Figure 4a). Negative lags indicate that Eurasian surface air temperature changes lead stratospheric temperature variability. While there is a significant correlation between the two at all lags such that cold Eurasian temperature occur in tandem with higher polar cap heights (i.e., a weaker vortex), the correlation is stronger for positive lags, demonstrating that the stronger direction of association is for higher polar cap heights leading to cold Eurasian surface conditions. Similar results are present for ERA-I data as well (supporting information Figure S3). Although these results do not necessarily suggest that Eurasian cooling was caused exclusively by stratospheric variability as there is upward coupling, it is clear that the former was enhanced by the latter.

The importance of stratosphere-to-surface downward coupling is further demonstrated by considering the stratospheric contribution to Eurasian wintertime "cold snaps." We start with every day in which Eurasian surface air temperature anomalies in DJF dropped below $-1^\circ$C. Each "event" is defined as the first day in which temperature anomalies dropped below this threshold, and events must be separated by at least 3 days in which temperature anomalies were warmer than this threshold. We then show the evolution of Eurasian surface air temperatures as a function of day after the start of the event in Figure 4b. We examine separately events in which the vortex was stronger as opposed to weaker than average on the day the cold snap started (blue and red lines in Figure 4b). There are more cold snaps with a weak vortex than with a strong vortex at the start (48 versus 30 events), and this effect is statistically significant at the 5% level as given by a binomial sign test. Though the intensities of the cold snaps are indistinguishable at the start of the cold snap by construction,
the cold snaps that start during weak vortex conditions persist for much longer — more than 60 days — while the cold snaps that start during strong vortex conditions dissipate in less than 15 days. The difference in Eurasian surface air temperature between the two is statistically significant at lags between 15 and 30 days and 55 to 60 days at the 5% level by a Student’s t test. We have computed the seasonal average surrounding the cold snaps, here defined as the average temperature for all lags shown on Figure 4b (10 days before to 60 days after), and cold snaps that start during a strong vortex are associated with near-climatological seasonal mean Eurasian surface air temperature (indicated on Figure). In contrast, cold snaps that start during a weak vortex are associated with cold conditions throughout the season and hence impact interannual and decadal variability.

Overall, we conclude that downward coupling from the stratosphere to the surface is the dominant direction of causation (though upward coupling does exist) and that cold snaps are more common and persist for longer if the vortex is weak and affect interannual variability of Eurasian temperatures only if the vortex is weak at the start of the event. While sorting through the many proposed mechanisms for the downward coupling is a topic of research [e.g., Garfinkel et al., 2013; Kidston et al., 2015], stratospheric anomalies couple down to the surface independent of any prior memory in the troposphere and extend the timescale over which surface climate variability is predictable [Charlton et al., 2004; Gerber et al., 2009; Sigmond et al., 2013; Tripathi et al., 2015; Scaife et al., 2016]. Combined with the results presented in Figures 1 and 2, it is evident that the expected Eurasian warming due to the increasing greenhouse gas concentrations was largely canceled by the stratospheric polar vortex weakening.

Finally, we consider the cause of these stratospheric changes. Did any forcing drive the observed weak vortices in the 2000s or was it internal variability? To answer this, we examine the ensemble mean response, as the ensemble mean is the model’s best estimate of the response to the time-evolving forcings included in all experiments such as greenhouse gases, sea surface temperatures, and sea ice. The ensemble mean of the GEOSCCM ensemble does not indicate weakening of the vortex in the 2000s despite time-evolving greenhouse gases, sea surface temperatures, and sea ice, though unforced internal variability is sufficiently strong that three members indicate decadal variability similar to that observed. Hence, the hiatus in these three model simulations was largely due to unforced variability (see supporting information Text S1 for additional discussion of the possible roles of sea ice, Eurasian snow cover changes, and SST changes) [e.g., Hardiman et al., 2008; Kim et al., 2014; Furtado et al., 2015; Garfinkel et al., 2015; Nakamura et al., 2016]. Stated another way, there is substantial internal variability in the polar lower stratosphere, and the apparent trend since 1990 in Eurasian surface temperatures may be just a byproduct of one specific manifestation of this chaotic decadal variability.

The above results suggest that the weakening of the polar vortex in the recent two decades, which is likely unforced decadal variability and may have been partly seeded by the Eurasian cooling itself, has significantly enhanced the recent Eurasian winter cooling. Three of the 29 GEOSCCM simulations capture wintertime Eurasian cooling in the 2000s comparable to that seen in the reanalyses data, and these three also simulate a weakened vortex as seen in the reanalyses data. As more than 10% of our model ensemble captured the hiatus (Figure 2a), we suggest that such hiatus events are not particularly rare and may recur in the future even as greenhouse gas concentrations rise. It is important to note that models lacking realistic stratospheric variability (i.e., with a low top) may underestimate the likelihood of such events, and hence, output from these models should be interpreted with caution. Conversely, models with a high top can simulate enhanced decadal-scale warming over Eurasia as compared to other regions if internal (unforced) variability results in a decadal-scale strengthening of the stratospheric polar vortex (e.g., Figure 2).

References
Baldwin, M. P., and D. W. J. Thompson (2009), A critical comparison of stratosphere-troposphere coupling indices, Q. J. R. Meteorol. Soc., 135, 1661–1672.
Baldwin, M. P., D. B. Stephenson, D. W. J. Thompson, T. J. Dunkerton, A. J. Charlton, and A. O’Neill (2003), Stratospheric memory and skill of extended-range weather forecasts, Science, 301, 636–640, doi:10.1126/science.1087143.
Bretherton, C. S., M. Widmann, V. P. Dymnikov, J. M. Wallace, and I. Blade (1999), The effective number of spatial degrees of freedom of a time-varying field, J. Clim., 12, 1990–2009.
Butler, A. H., D. J. Seidel, S. C. Hardiman, N. Butchart, T. Birner, and A. Match (2015), Defining sudden stratospheric warmings, Bull. Am. Meteorol. Soc., 96(11), 1913–1928.
Charlton, A. J., A. O’Neill, W. Lahoz, and A. Massacand (2004), Sensitivity of tropospheric forecasts to stratospheric initial conditions, Q. J. R. Meteorol. Soc., 130(600), 1771–1792.
Garfinkel, C. I., and N. Harnik (2016), The non-Gaussianity and spatial asymmetry of temperature extremes relative to the storm track: The role of horizontal advection, J. Clim., 30(2), doi:10.1175/JCLI-D-15-0806.1.

Garfinkel, C. I., D. W. Waugh, and E. P. Gerber (2013), The effect of tropospheric jet latitude on coupling between the stratospheric polar vortex and the troposphere, J. Clim., 26(6), 2077–2095, doi:10.1175/JCLI-D-12-00301.1.

Garfinkel, C. I., M. M. Hurwitz, and L. D. Oman (2015), Effect of recent sea surface temperature trends on the Arctic stratospheric vortex, J. Geophys. Res. Atmos., 120, 5404–5416, doi:10.1002/2015JD023284.

Gerber, E., C. Orbe, and L. M. Polvani (2009), Stratospheric influence on the tropospheric circulation revealed by idealized ensemble forecasts, Geophys. Res. Lett., 36, L24801, doi:10.1029/2009GL040913.

Hardiman, S. C., P. J. Kushner, and J. Cohen (2008), Investigating the ability of general circulation models to capture the effects of Eurasian snow cover on winter climate, J. Geophys. Res., 113, D21123, doi:10.1029/2008JD010623.

Kidston, J., A. A. Scaife, S. C. Hardiman, D. M. Mitchell, N. Butchart, M. P. Baldwin, and L. J. Gray (2015), Stratospheric influence on tropospheric jet streams, storm tracks and surface weather, Nat. Geosci., 8(6), 433–440.

Kim, B.-M., S.-W. Son, S.-K. Min, J.-H. Jeong, S.-J. Kim, X. Zhang, T. Shim, and J.-H. Yoon (2014), Weakening of the stratospheric polar vortex by Arctic sea-ice loss, Nat. Commun., 5, doi:10.1038/ncomms6566.

Kobayashi, S., et al. (2015), The JRA-55 reanalysis: General specifications and basic characteristics, J. Meteorol. Soc. Jpn., 93(1), 5–48.

Kolstad, E. W., T. Breteig, and A. A. Scaife (2010), The association between stratospheric weak polar vortex events and cold air outbreaks in the Northern Hemisphere, Q. J. R. Meteorol. Soc., 136(649), 886–893.

Kosaka, Y., and S.-P. Xie (2013), Recent global-warming hiatus tied to equatorial Pacific surface cooling, Nature, 501(7467), 403–407, doi:10.1038/nature12534.

Kug, J.-S., J.-H. Jeong, Y.-S. Jang, B.-M. Kim, C. K. Folland, S.-K. Min, and J.-S. Kim (2015), Two distinct influences of arctic warming on cold winters over North America and East Asia, Nat. Geosci., 8, 759–762, doi:10.1038/ngeo2517.

Lehtonen, I., and A. Y. Karpechko (2016), Observed and modeled tropospheric cold anomalies associated with sudden stratospheric warmings, J. Geophys. Res. Atmos., 121, 1591–1610, doi:10.1002/2015JD023860.

Li, C. B., Stevens, and J. Marotzke (2015), Eurasian winter cooling in the warming hiatus of 1998–2012, Geophys. Res. Lett., 42, 8131–8139, doi:10.1002/2015GL065327.

Manzini, E., et al. (2014), Northern winter climate change: Assessment of uncertainty in CMIP5 projections related to stratosphere-troposphere coupling, J. Geophys. Res. Atmos., 119, 7979–7998, doi:10.1002/2013JD021403.

Meeth, G. A., and H. Teng (2014), CMIP5 multi-model hindcasts for the mid-1970s shift and early 2000s hiatus and predictions for 2016–2035, Geophys. Res. Lett., 41, 1711–1716, doi:10.1002/2014GL065926.

Meinshausen, M., et al. (2011), The RCP greenhouse gas concentrations and their extensions from 1765 to 2300, Clim. change, 109(1–2), 213–241.

Mołod, A., L. Takacs, M. Suarez, J. Bacmeister, J.-S. Song, and A. Eichmann (2012), The GEOS-5 atmospheric general circulation model: Mean climate and development from MERRA to Fortuna, Tech. Rep. Ser. on Global Model. and Data Assimilation 28, Natl. Aeronautics and Space Admin., Goddard Space Flight Cent., Greenbelt, Md.

Nakamura, T., K. Yamazaki, K. Iwamoto, M. Honda, Y. Miyoshi, Y. Ogawa, Y. Tomikawa, and J. Ukiti (2016), The stratospheric pathway for arctic impacts on midlatitude climate, Geophys. Res. Lett., 43, 3494–3501, doi:10.1002/2016GL068330.

Pawson, S., R. Stolarski, A. R. Douglass, P. A. Newman, J. E. Nielsen, S. M. Frith, and M. L. Gupta (2008), Goddard Earth Observing System chemistry-climate model simulations of stratospheric ozone-temperature coupling between 1950 and 2005, J. Geophys. Res., 113, D24103, doi:10.1029/2007JD009511.

Pyper, B. J., and R. M. Peterman (1998), Comparison of methods to account for autocorrelation in correlation analyses of fish data, Can. J. Fish. Aquat. Sci., 55(9), 2127–2140, doi:10.1139/f98-104.

Rayner, N., P. Brohan, D. Parker, C. Folland, J. Kennedy, M. Vainike, T. Ansell, and S. Tett (2006), Improved analyses of changes and uncertainties in sea surface temperature measured in situ since the mid-nineteenth century: The HadSST2 dataset, J. Clim., 19(3), 446–469, doi:10.1175/JCLI3637.1.

Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Wang (2002), An improved in situ and satellite SST analysis for climate, J. Clim., 15(13), 1609–1625, doi:10.1175/1520-0442(2002)015<1609:AIISAS>2.0.CO;2.

Rienecker, M. M., et al. (2011), MERRA: NASA’s Modern-Era Retrospective Analysis for Research and Applications, J. Clim., 24, 3624–3648, doi:10.1175/JCLI-D-11-00051.1.

Rienecker, M. M., et al. (2008), The GEOS-5 data assimilation system—Documentation of versions 5.0.1, 5.1.0, and 5.2.0, Tech. Rep. Ser. on Global Model. and Data Assimilation 27, Natl. Aeronaut. and Space Admin., Goddard Space Flight Cent., Greenbelt, Md.

Santer, B. D., et al. (2008), Consistency of modelled and observed temperature trends in the tropical troposphere, Int. J. Climatol., 28(13), 1703–1722.

Scaife, A., et al. (2016), Seasonal winter forecasts and the stratosphere, Atmos. Sci. Lett., 17(1), 51–56.

Sigmond, M., J. Scinocca, V. Kharin, and T. Shepherd (2013), Enhanced seasonal forecast skill following stratospheric sudden warmings, Nat. Geosci., 6(2), 98–102, doi:10.1038/ngeo1698.

Stocker, T., et al. (2013), IPCC 2013: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge Univ. Press, Cambridge, U. K., and New York.
Thompson, D. W. J., M. P. Baldwin, and J. M. Wallace (2002), Stratospheric connection to northern Hemisphere wintertime weather: Implications for prediction, J. Clim., 15, 1421–1428, doi:10.1175/1520-0442(2002)015.

Thorne, P., S. Outten, I. Bethke, and O. Seland (2015), Investigating the recent apparent hiatus in surface temperature increases: 2. Comparison of model ensembles to observational estimates, J. Geophys. Res. Atmos., 120, 8597–8620, doi:10.1002/2014JD022805.

Tomassini, L., E. P. Gerber, M. P. Baldwin, F. Bunzel, and M. Giorgetta (2012), The role of stratosphere-troposphere coupling in the occurrence of extreme winter cold spells over northern Europe, J. Adv. Model. Earth Syst., 4, M00A03, doi:10.1029/2012MS000177.

Trenberth, K. E., and J. T. Fasullo (2013), An apparent hiatus in global warming?, Earth’s Future, 1(1), 19–32, doi:10.1002/2013EF000165.

Trenberth, K. E., J. T. Fasullo, G. Branstator, and A. S. Phillips (2014), Seasonal aspects of the recent pause in surface warming, Nat. Clim. Change, 4(10), 911–916, doi:10.1038/nclimate2341.

Tripathi, O. P., et al. (2015), The predictability of the extra-tropical stratosphere on monthly timescales and its impact on the skill of tropospheric forecasts, Q. J. R. Meteorol. Soc., 141, 987–1003, doi:10.1002/qj.2432.