Global warming in an independent record of the past 130 years

D. M. Anderson,[1] E. M. Mauk,[1,2] E. R. Wahl,[1] C. Morrill,[1,3] A. J. Wagner,[1,3]
D. Easterling,[4] and T. Rutishauser[5]

Received 18 October 2012; revised 20 November 2012; accepted 21 November 2012; published 16 January 2013.

[1] The thermometer-based global surface temperature time series (GST) commands a prominent role in the evidence for global warming, yet this record has considerable uncertainty. An independent record with better geographic coverage would be valuable in understanding recent change in the context of natural variability. We compiled the Paleo Index (*PI*) from 173 temperature-sensitive proxy time series (corals, ice cores, speleothems, lake and ocean sediments, historical documents). Each series was normalized to produce index values of change relative to a 1901–2000 base period; the index values were then averaged. From 1880 to 1995, the index trends significantly upward, similar to the GST. Smaller-scale aspects of the GST including two warming trends and a warm interval during the 1940s are also observed in the *PI*. The *PI* extends to 1730 with 67 records. The upward trend appears to begin in the early 19th century but the year-to-year variability is large and the 1730–1929 trend is small. **Citation:** Anderson, D. M., E. M. Mauk, E. R. Wahl, C. Morrill, A. J. Wagner, D. Easterling, and T. Rutishauser (2013), Global warming in an independent record of the past 130 years. *Geophys. Res. Lett.*, *40*, 189–193, doi:10.1029/2012GL054271.

1. Introduction

[2] The time series of global average surface temperature (GST) is useful in understanding natural variability and anthropogenic warming during the instrumental period. Three temperature series [*Brohan et al.*, 2006; *Hansen et al.*, 2010; *Smith et al.*, 2008] have been produced, differing in their data selection, processing, and bias corrections but using mostly the same underlying temperature measurements. Beginning in 1880, these temperature anomaly series show an initial period of little change, warming from the 1910s to the 1940s, a peak in the 1940s, and a second period of warming from the 1960s to the present. The GST has increased by 0.6–0.7°C over the past century [*IPCC*, 2001; *Solomon et al.*, 2007], and the 2007 IPCC Fourth Assessment Report notes that the linear warming trend for the last 50 years (0.13°C/decade) is twice that for the last 100 years.

[3] The GST and its corrections and potential biases have been intensely scrutinized [*Hansen et al.*, 2010]. The number of observations has changed through time, instruments have changed, and stations have been moved—including from more dense portions of urban areas to relatively open airport locations. Land use/land cover characteristics around the stations have changed, with urbanization and large-scale irrigation being two obvious differences. Similarly for sea-surface temperatures, instruments and procedures have changed. For example early ship-based observations utilized buckets while later observations measured water intake temperatures [*Kennedy et al.*, 2011; *Rayner et al.*, 2006; *Smith et al.*, 2008]. Many of the potential biases and problems have been accounted for through data corrections; however, these adjustments are themselves uncertain. To test for positive trend, determine whether the trend is constant or accelerating, globally pervasive or regionally distinct, neither calibration of proxy records to temperature, nor annual resolution is required. The main goal of this paper is to test for the presence of large-scale, multidecadal trends in temperature-sensitive proxy evidence during the instrumental period 1880 to present. A secondary goal is to assess the trend prior to 1880.

2. Methods

[4] We selected paleoclimatic ("paleo") proxy time series that contained at least 10 samples between 1880 and 1995 to achieve decadal resolution. The samples are mostly evenly distributed across the 116 years and 173 sites; 40 records are decadally resolved and the remaining 133 are annually resolved. Data gaps (4%) near the beginning (1880) arise because corals are short-lived and many ice cores are short. Data gaps near the end (1995) arise because many paleo records were collected decades ago (individual series listed in the Supporting Information). We included proxies for which sensitivity to temperature is attributed to a physical link, accepting that some of these proxies are also influenced by other environmental variables. Examples of physical linkages include the oxygen isotope ratio in biogenic carbonates (corals, foraminifers), which decreases by approximately 0.25 per mil per °C [*Epstein et al.*, 1953]. Sr/Ca decreases in corals by 0.06 mmol/mol per °C [*Correge*, 2006]. The oxygen isotopes of water that accumulates in glaciers and ice sheets vary with temperature; one of the original global, geographically based estimates was an increase of 0.7 per mil per °C [*Dansgaard*, 1964]. The significance of recent trends in compilations of ice core series has been described for Antarctic [*Schneider and Steig*, 2008] as well as tropical locations [*Thompson et al.*, 2003]. Sixty

This article was published online on 16 January 2013. Subsequently, some errors were identified, and the correction was published on 28 January 2013.

All Supporting Information may be found in the online version of this article.

[1]NOAA's National Climatic Data Center, Boulder, Colorado, USA.
[2]Environment and Sustainability Program, School of Earth, Ocean, and Environment, University of South Carolina, Columbia, South Carolina, USA.
[3]Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, Colorado, USA.
[4]NOAA's National Climatic Data Center, Asheville, North Carolina, USA.
[5]Oeschger Centre for Climate Change Research and Institute of Geography, University of Bern, Bern, Switzerland.

Corresponding author: D. M. Anderson, NOAA's National Climatic Data Center, Boulder, CO, USA. (david.m.anderson@noaa.gov)

0094-8276/13/2012GL054271

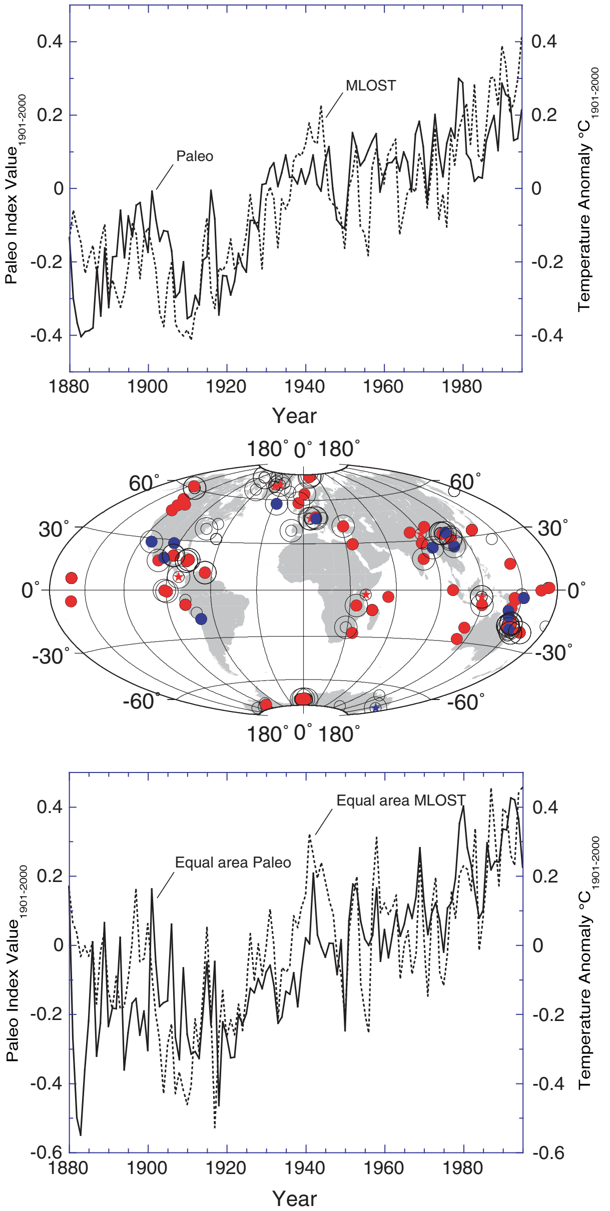

Figure 1. (top) Paleo Index (solid) and the merged land-ocean surface temperature anomalies (MLOST, dashed) relative to 1901–2000. The range of the paleo trends index values is coincidentally nearly the same as the GST although the quantities are different (index values versus temperature anomalies °C). (middle) Sites with significant warming trend 1880–1995 (red, $p < 0.05$), less significant trend (red star, $0.05 < p < 0.20$), and no significant trend (open circle, $p > 0.20$); also sites with a significant cooling trend (blue, $p < 0.05$) or less significant cooling trend (blue star, $0.05 < p < 0.20$). Double circles show 67 sites extending to 1730; Common-spatial-coverage comparison shows *PI* and MLOST area-weighted global averages where only 10° × 10° cells containing both *PI* and MLOST observations were included in the average.

percent of the 170 records are oxygen isotopes preserved in ice or carbonate, where the relative influence of source, trajectory, amount, and temperature is uncertain and time-variable. We hypothesize that the confounding influences cancel in the global average, and test by comparing the full series with a subset that excludes tropical oxygen isotope series. To test for the presence of a regional or proxy-specific, nontemperature influence, we mapped the location of significant trend, and we analyzed the trend in subsets of the data that exclude potentially biased proxies. Forty biological, as opposed to physical proxies were included [*Lough and Barnes*, 1997; *Rutishauser et al.*, 2007; *Rutishauser et al.*, 2008]. To avoid potential loss of low-frequency variations, we excluded tree ring series. Tree ring growth-width data are routinely detrended, potentially removing low frequency variance depending on the detrending method used [*Cook et al.*, 1995]. To remain independent of the instrumental record, we used the raw data rather than reconstructed temperatures. Each series was normalized by subtracting the 1901–2000 mean and then dividing by the standard deviation. Yearly values were calculated for each of the $m = 173$ series (less-than-annually resolved series have years with missing values), proxy series whose values decrease with warming (Sr/Ca, $\delta^{18}O_{carbonate}$) were multiplied by –1, and then the individual series were averaged to form the index. The number of proxy measurements decreases rapidly between 1995 and 1999, and prior to 1730 and after 1999 less than 66 series contain observations. We thus chose 1730–1995 as the period of the Paleo Index (*PI*). We did not select or reject series based on the strength of the time-series correlation with an overlapping temperature record, unlike the procedure used in *Mann et al.* [2008, 2009] and *Wilson et al.* [2006]; this would necessarily include records containing trend. Some of the Mann et al. and Wilson et al. data were included because of the physical link described above, but other records from Mann et al. were excluded (for example Lake Pallcacocha sediment color, Chesapeake salinity, lake Malawi biogenic silica %) because they lacked a physical link or were tree ring-width data. Our compilation is extensive but not exhaustive. It aggregates and builds on previous compilations [*Ljungqvist et al.*, 2012; *Lough and Barnes*, 1997; *Schneider and Steig*, 2008; *Wilson et al.*, 2006], but other data may exist that could be added later. We used the age determinations as originally reported. The reported age uncertainty varies, and is lowest in the layer-counted proxies and largest for the proxies that relied on radiocarbon dating. For the layer-counted records, comprising most of the series, we estimate the age uncertainty to be 5%, thus an uncertainty of +/– 1 years in 1995 increases to +/– 6 years in 1880 and increases to +/– 13 years in 1730 (discussed below). Forty series have decadal (as opposed to annual) resolution, and these series also have the greatest age uncertainty. All selected records are subject to multiple environmental influences, and a key assumption of our approach, as well as that of Mann et al., and to a lesser extent the compilers of the GST, is that nontemperature influences are minimized by averaging many observations together.

3. Results

[5] We compared the *PI* with the merged land-ocean surface temperature (MLOST) anomalies [*Smith et al.*, 2008] developed by the National Climatic Data Center (Figure 1). Both series were adjusted to have zero mean for the period 1901–2000. Both series contain a warming trend. The trend in the *PI* (slope = 0.043 ± 0.011 s.d. units per decade) between 1880 and 1995 is highly significant at the $\ll 0.01$ level ($t^* = 7.90$; autocorrelation-adjusted $df = 30$, based on

Table 1. Trend (s.d. Units/Decade) and 95% Adjusted Confidence Interval, Number of Series (m) Comprising the Global Average at Start of Interval, and Probability (p) the Estimated Trend Value (or Higher) Would Be Expected to Occur if the True Trend Value is Zero

Series	Trend	m	p
Marine only$_{1880–1995}$	0.044 ± 0.010	91	8.80E-12
No coral extension$_{1880–1995}$	0.051 ± 0.017	136	2.76E-06
No tropical $\delta^{18}O_{1880–1995}$	0.031 ± 0.011	139	2.46E-06
Frozen grid$_{1880–1995}$	0.031 ± 0.011	67	9.53E-07
Common spatial coverage$_{1880–1995}$	0.048 ± 0.010	13-50	1.42E-13
$PI_{1880–1995}$	0.043 ± 0.011	173	1.31E-08
$PI_{1980–1995}$	0.071 ± 0.152	173	0.30
$PI_{1920–1935}$	0.260 ± 0.056	173	3.34E-07
$PI_{1730–1928}$	0.016 ± 0.005	67	1.21E-09
$PI_{1929–1995}$	0.017 ± 0.015	173	1.53E-03

116 years; $p = 1.31e\text{-}8$). The t-test for nonzero trend ($t_b = b/s_b$, where b is the estimated trend and s_b its estimated standard error) was conservatively adjusted for autocorrelation by reducing the effective sample size when calculating s_b and also in the determination of the critical t value [*Santer et al.*, 2000]. This adjustment was made to both the global *PI* and to 133 individual series (map, Figure 1) where the autocorrelation could be assessed; the t-statistic for the remaining time series was calculated without adjustment. The significance of the *PI* trend was further evaluated using a bootstrapping approach based on resampling the trend regression residuals [*Li et al.*, 2007]. For 1000 iterations, the median of the estimated trend is 0.042 (ranging from 0.029 (2.5% quantile) to 0.052 (97.5% quantile), with none of the bootstrap iterations giving a slope value near to or less than zero), confirming the trend's significance. In addition to the shared long-term upward trend, many of the smaller-scale features of the instrumental series appear in the *PI*, including two warming trends and a warm interval during the 1940s, leading to a correlation of 0.76 ($n = 116$) between the *PI* and the GST. At the finest scale (0.1 s.d. and 0.1°C) the two records diverge. Some of the fine-scale differences may be attributed to age uncertainty. Thus, using only temperature-sensitive paleo proxy records uncalibrated to instrumental data, it is possible to conclude that the warming trend in the GST is supported by independent evidence.

[6] The increased rate of warming in the GST in the late 20th century has been cited as another diagnostic of anthropogenic change. In contrast, natural forcing (volcanic, solar) is not accelerating [*Crowley*, 2000]. Changes in the trend were detected using two-phase linear regression [*Lund and Reeves*, 2002] to test the hypothesis that one regression line fits the 1880–1995 time series better than two or more, accounting for the different degrees of freedom. According to this analysis the *PI* trend changes in 1928 (0.05 significance level). During the last 15 years (1980–1995) the *PI* trend is 65% greater (0.071) than the long-term trend (0.043), but not unprecedented. The upward trend from 1920–1935 was larger (0.260 ± 0.056) and (Table 1, $PI_{1920–1935}$). Among the 81 sites that trend significantly upward ($p < 0.05$), the trend increases after 1980 at 37 sites. Sixty percent of the coral-geochemistry proxy records and 25% percent of the ice core records exhibit an increasing post-1980 trend.

[7] We additionally examined the spatial pattern in the trend and searched for potential biases among proxy subsets, recognizing that some regional or proxy-specific characteristics are lost in the averaging procedure. The significant positive (warming) trend is widespread (Figure 1), found in Arctic, subpolar, temperate, and tropical sites. Thirteen sites have significant negative (cooling) trend ($p < 0.05$; t^* and df were adjusted for autocorrelation and reported for each series in the supplemental information), which is consistent with the instrumental record in which not all sites have the same sign of change as the global average. A pattern analysis [*Rowlingson and Diggle*, 1993] confirms that sites with trend are not confined to specific regions such as the Great Barrier Reef corals or Greenland ice cores. Greenland ice core oxygen isotopes lack significant trend, unlike other circum-North Atlantic proxies. Some Antarctic ice cores (including δD and δ^{18}O) contain significant trends, while some do not. In contrast to the instrumental record, where the density of observations is greatest over Europe and North America, the paleo sites are more evenly distributed (Figure 1). Therefore, the distribution of the paleo sites supports the global extent of warming observed in the instrumental record. If the warming trend were localized to the midlatitude Northern Hemisphere, it would not likely be evident in this compilation of paleo sites. There are only 30 sites between 30°N and 60°N (Figure 1).

[8] Uneven spatial coverage or over-representation of a group of samples (e.g., Great Barrier Reef corals, ITASE (International Trans Antarctic Scientific Expedition) ice cores), along with changing sample size through time, might bias the global trend or the comparison with MLOST. To test for and remove this potential for bias, trends were calculated on subsets of the data, and an additional, common-spatial-coverage estimate was made. We made subsets (Table 1) excluding terrestrial data (marine only, number of sites $m = 91$), excluding coral extension rate proxies ($m = 136$), excluding all tropical (latitude <23°) δ^{18}O ($m = 139$), and averaging only sites existing in 1730 ($m = 67$, "frozen grid"). The trend in the $PI_{\text{frozen grid } 1880–1995}$ (0.031) is the same as the *PI* trend within the 0.05 CI Confidence Interval and highly significantly positive, and we conclude that the effect of changing sample density is insignificant (Table 1). The marine-only trend (0.044) is nearly identical to the *PI* trend, which counters criticism of marine observations; proxy data indicate that the ocean surface is warming as much as the land. When tropical δ^{18}O records are removed to eliminate possible positive-trend bias from increasing rainfall (light δ^{18}O) the trend decreases to 0.031 (not different from 0.043 ± 0.011). In each withholding-data subset the trend remains highly significant ($p \ll 0.01$)(Table 1). To compare the *PI* with MLOST on the basis of common spatial coverage, we formed an area-weighted global average from 10° × 10° squares, averaging all time series found within a square and including on a year-by-year basis only squares containing at least one paleo and at least one instrumental observation. The result (Figure 1) is noisy, containing more year-to-year change, but retains the overall range and upward trend observed in the original *PI* and MLOST time series. Averaging by 10° squares reduces the number of Great Barrier Reef extension rate and geochemical proxy series that contribute to the index, and likewise limits the contribution from Antarctic ITASE and other groups of records. Sites with two proxies (e.g., Mg/Ca and δ^{18}O) contributed only one (averaged) series in this weighting. Restricting the average to squares with both *PI* and MLOST data means that fewer time series comprise the global mean, 30–50 series between 1904–1994, and 13–30 series

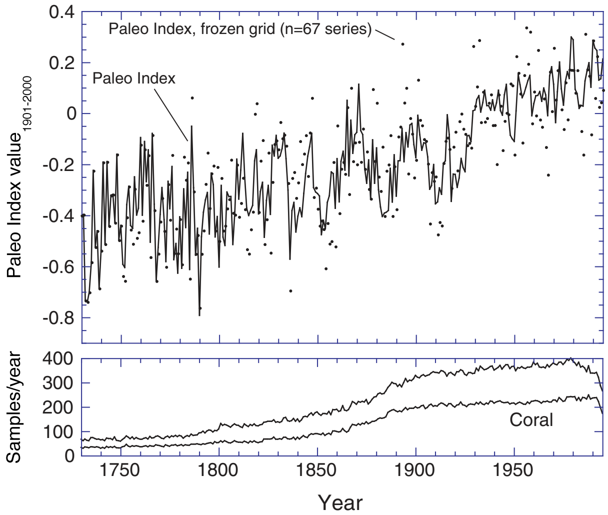

Figure 2. (top) Paleo index extending back to 1730 (solid line), the Paleo index derived by averaging only the sites that begin prior to 1730 (dots). (bottom) Total number of observations per year and the number of coral observations per year.

prior to 1903 when the decrease in squares containing instrumental observations is substantial. Despite severe reductions in the number of sites forming the global average, the *PI* and MLOST data retain their significant trend, their similarity to one another (Figure 1), and their similarity to the original series (Figure 1).

[9] The *PI* can be extended back to 1730 (Figure 2). Similar to longer temperature-calibrated multiproxy reconstructions, the PI_{extended} trends upwards a century prior to the start of upward trend in the MLOST [*Hegerl et al.*, 2007; *Mann et al.*, 2008, 2009; *Moberg et al.*, 2005]. The coldest temperatures in the multiproxy reconstructions occurred between 1600 and 1700 A.D., and fingerprint-based detection and attribution has linked European warming prior to 1880 to a combination of greenhouse, solar, and volcanic forcing [*Hegerl et al.*, 2011]. Early trends are small, and the trend from 1730 to 1928 (when a change point is detected) is 0.015 ± 0.005. The time series utilizing the 67 sites existing in 1730 ($PI_{\mathrm{frozen\ grid}}$) (Figure 2, dots) is initially identical to the PI_{extended}; however, the year-to-year variability in the $PI_{\mathrm{frozen\ grid}}$ increases after 1800 with respect to the PI_{extended}, which is attributed to the increased site density reducing noise in the PI_{extended}. There are limitations to the interpretation of the early period. Many of the well-dated coral records drop out between 1900 and 1800. Age uncertainty for the layer-counted records increases 5% per year (+/−10 years at 1800) (however, phenology records are extremely well dated at the annual time scale). Coverage remains global but the number of ice cores (from more extreme environments) relative to corals increases. The remarkable aspect prior to 1880 is the persistent cooler than average index values 1730–1900. Other aspects, including the significance of the early trend, the magnitude of decadal variability, and the onset of warming, will require an improved data set.

4. Summary

[10] The global-average time series compiled from 170 temperature-sensitive paleo proxies indicates a significant warming trend from 1880–1995. Derived from multiple proxies with global distribution, the *PI* provides independent evidence of the warming observed in the thermometer-based record. The *PI* extends to 1730 with 67 records. Warming appears to begin around 1800 based on the 67 series that extend back to 1730; index values are well below the 1901–2000 mean from 1730 into the 20th century.

[11] **Acknowledgments.** Elyse M. Mauk acknowledges the University of South Carolina MEERM-AWNES graduate program and Jennifer R. Pournelle for providing specific mentorship and endless patience, and the University of Colorado at Boulder Museum of Natural History with Jaelyn J. Eberle, whose support and influence helped facilitate the project and fostered collaboration.

References

Brohan, P., et al. (2006), Uncertainty estimates in regional and global observed temperature changes: a new dataset from 1850, *J. Geophys. Res.*, *111*, D12106.

Cook, E. R., et al. (1995), The 'segment length curse' in long tree-ring chronology development for palaeoclimatic studies, *The Holocene*, *5*, 229–237.

Correge, T. (2006), Sea surface temperature and salinity reconstruction from coral geochemical tracers, *Palaeogeogr. Palaeocl.*, *232*(2-4), 408–428.

Crowley, T. J. (2000), Causes of climate change over the past 1000 years, *Science*, *289*, 270–277.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*, 436–468.

Epstein, S., et al. (1953), Revised carbonate-water isotopic temperature scale, *B. Geol. Soc. Am.*, *64*, 1315–1326.

Hansen, J., et al. (2010), Global surface temperature change, *Rev. Geophys.*, *48*, RG4004.

Hegerl, G. C., et al. (2007), Detection of human influence on a new, validated, 1500-year temperature reconstruction, *J. Clim.*, *20*, 650–666.

Hegerl, G. C., et al. (2011), Influence of human and natural forcing on European seasonal temperatures, *Nat. Geosci.*, *4*, 99–103.

IPCC (2001), Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge.

Kennedy, J. J., et al. (2011), Reassessing biases and other uncertainties in sea surface temperature observations measured in situ since 1850: 2. Biases and homogenization, *J. Geophys. Res.*, *116*, 1–22.

Li, B., et al. (2007), The "Hockey Stick" and the 1990s: A statistical perspective on reconstructing hemispheric temperatures, *Tellus*, *59*, 591–598.

Ljungqvist, F. C., et al. (2012), Northern hemisphere temperature patterns in the last 12 centuries, *Climates of the Past*, *8*, 227–249.

Lough, J. M., and D. J. Barnes (1997), Several centuries of variation in skeletal extension, density and calcification in massive Porites colonies from the Great Barrier Reef: a proxy for seawater temperature and a background of variability against which to identify unnatural change, *J. Exp. Mar. Biol. Ecol. 211*, 29–67.

Lund, R., and J. Reeves (2002), Detection of Undocumented Changepoints: a Revision of the Two-Phase Regression Model, *J. Clim.*, *15*, 2547–2554.

Mann, M. E., et al. (2008), Proxy-based reconstructions of hemispheric and global surface temperature variations over the past two millennia, *Proc. Nat. Acad. Sci. U.S.A.*, *105*(36), 13,252–13,257.

Mann, M. E., et al. (2009), Global signatures and dynamical origins of the Little Ice Age and Medieval Climate Anomaly, *Science*, *326*, 1256–1260.

Moberg, A., et al. (2005), Highly variable Northern Hemisphere temperatures reconstructed from low- and high-resolution proxy data, *Nature*, *433*, 613–617.

Rayner, N. A., et al. (2006), Improved analyses of changes and uncertainties in marine temperature measured in situ since the mid-nineteenth century: the HadSST2 dataset, *J. Clim.*, *19*, 446–469.

Rowlingson, B. S., and P. J. Diggle (1993), SPLANCS: Spatial point pattern analysis code in S-Plus, *Comput. Geosci.*, *19*(5), 627–655.

Rutishauser, T., et al. (2007), A phenology-based reconstruction of interannual changes in past spring seasons, *J. Geophys. Res.*, *112*, G04016.

Rutishauser, T., et al. (2008), Swiss Spring Plant Phenology 2007: Extremes, a multi-century perspective and changes in temperature sensitivity, *Geophys. Res. Lett.*, *35*(L05703).

Santer, B. D., et al. (2000), Statistical significance of trends and trend differences in layer-average atmospheric temperature time series, *J. Geophys. Res.*, *105*(D6), 7337–7356.

Schneider, D. P., and E. J. Steig (2008), Ice cores record significant 1940s Antarctic warmth related to tropical climate variability, *Proc. Nat. Acad. Sci. U.S.A.*, *105*(34), 12,154–12,158.

Smith, T. M., et al. (2008), Improvements to NOAA's historical merged land-ocean surface temperature analysis (1880-2006), *J. Clim.*, *21*, 2283–2296.

Solomon, S., et al. (2007), Climate Change 2007: The physical science basis. Contribution of working group 1 to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, New York.

Thompson, L. G., et al. (2003), Tropical glacier and ice core evidence of climate change on annual to millennial time scales, *Clim. Change*, *59*, 137–155.

Wilson, R., et al. (2006), Two-hundred-fifty years of reconstructed and modeled tropical temperatures, *J. Geophys. Res.*, *11*, 1–13.