A multi-century ice-core perspective on 20th-century climate change with new contributions from high-Arctic and Greenland (PARCA) cores

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ABSTRACT. A global collection of high temporally resolved ice-core-derived $\delta^{18}O$ records is examined to assess whether the proxy records are consistent with contemporaneous observed temperature variations in their respective regions. This is prerequisite to using the older parts of the proxy ($\delta^{18}O$) records to assess whether 20th-century temperatures remain within the range of longer-term natural variability. Excluding the high plateaus in East and West Antarctica where 20th-century temperatures show modest cooling, the ice-core records from other regions suggest modest to strong 20th-century warming. The recent warming over Greenland has been modest and spatially variable. The 20th-century warming over both the Barents Sea region and the Tibetan Plateau now falls well outside the range of prior longer-term temperature variability. Similarly, over the South American Andes and the Antarctic Peninsula the recent warming exceeds the long-term mean for the last 1000 and 500 years, respectively. The ice fields in these regions are in danger of being compromised or lost if the present warming trend in these regions persists.

INTRODUCTION

Since the late 1960s when the first long cores were recovered from Greenland and Antarctica, ice cores have provided unique details about the nature of the Earth's climate system with emphasis on glacial–interglacial cycles. Recently, greater attention has been focused on reconstructing high-resolution (optimally annual) proxy histories for the Holocene, with emphasis on the last few thousand years. This attention results, in part, from the need to assess the well-documented 20th-century warming of the Earth's surface temperatures from a longer-term perspective (Mann and others, 1999; Levitus and others, 2000; Hansen and others, 2001; Jones and Moberg, 2003; Moberg and others, 2005). Restricted to very cold and/or high locations, ice-core-derived proxy histories provide a partial picture of past climatic and environmental conditions. These frozen archives are powerful contributors to multi-proxy reconstructions, providing multiple lines of evidence (e.g. changes in atmospheric dustiness and chemistry, in mass accumulation and in temperature) from remote locations where other proxy data are unavailable.

Over the last few decades, the Ice Core Paleoclimate Research Group at The Ohio State University (OSU) has collected a suite of tropical and subtropical cores from the South American Andes and the Tibetan Plateau, as well as polar cores from Antarctica, Greenland and Franz Josef Land (FJL), Russian Arctic. The analyses of these ice cores have emphasized the highest time resolution possible. More recently, a number of new annually resolved ice-core records from Greenland have been added to the OSU collection. The cores were acquired between 1995 and 1999 as part of the NASA–NSF (US National Science Foundation) PARCA (Program for Arctic Regional Climate Assessment) project (Thomas and others, 2001). These cores are particularly valuable as they come from sites that are widely distributed along the western side of Greenland. Previously, most available high-resolution Greenland ice-core histories that extend back more than a century were drilled in central and north-central Greenland or in southern Greenland near the Dye 2 and Dye 3 radar stations (now abandoned).

ICE-CORE-DERIVED $\delta^{18}O$ RECORDS

Figure 1 presents the decadally averaged $\delta^{18}O$ records organized geographically (north to south), along with regional maps showing their locations. The records from Tibet, the Andes and Antarctica have been published separately, but not collectively with the newer $\delta^{18}O$ histories from Greenland and FJL. This synthesis is limited to the last millennium, as many of the records are relatively short, extending back just a few centuries to one half-millennium. Six of the $\delta^{18}O$ records extend to earlier millennia, four into the Last Glacial Stage. The longer records include those from the Dunde and Guliya ice cores from the Tibetan Plateau (Thompson and others, 1989, 1997, respectively), the Quelccaya and Huascarán cores from Peru (Thompson and others, 1986, 1995, respectively), the Sajama ice cap in Bolivia (Thompson and others, 1998) and the Plateau Remote site from East Antarctica (Mosley-Thompson, 1996). Table 1 provides basic information about each core. Details of the dating procedures are in the references.

The $\delta^{18}O$ histories from two different drilling projects at the Summit Site in central Greenland are included for comparison with the newer Greenland records (Fig. 1). The Greenland Ice Sheet Project 2 (GISP2) record (Grootes and others, 1993) is shown along with a composite from two

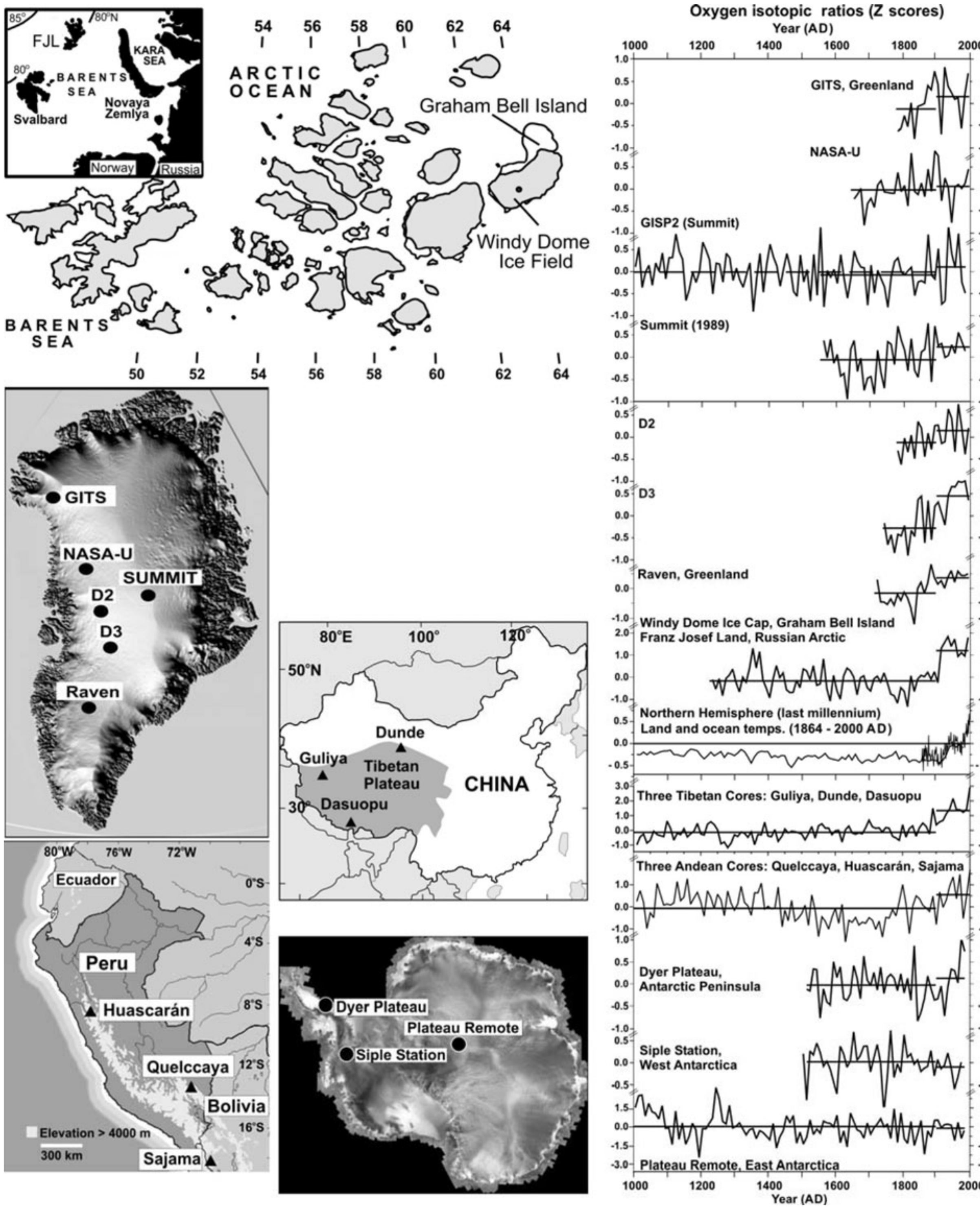

Fig. 1. Decadally averaged $\delta^{18}O$ histories are shown from north to south (top to bottom). Horizontal bars in each record show the average $\delta^{18}O$ values for the pre- and post-1900 time intervals. The 1000 year Northern Hemisphere reconstruction is from Mann and others (1999) and updated by Mann and Jones (2003). Superimposed is the observed near-surface temperature record (Jones and Moberg, 2003). Maps show locations for the cores.

cores, called Site T cores, drilled 4 km apart at the Summit in 1989 by OSU (Mosley-Thompson and others, 1993). The $\delta^{18}O$ history from the Windy Dome Ice Field on Graham Bell Island, FJL, (Henderson, 2002) is included along with the five new Greenland $\delta^{18}O$ histories: GITS, NASA-U, D2, D3, Raven (Fig. 1), collected by the PARCA initiative. With the exception of the GISP2 core, all records discussed here were analyzed and dated by the authors using consistent procedures. Moreover, unlike many of the earlier Greenland records, the timescales for the PARCA cores were constructed using three seasonally varying parameters: the insoluble dust concentrations and oxygen isotopic ratios ($\delta^{18}O$), both measured at OSU, and hydrogen peroxide (H_2O_2) measured at the University of Arizona (McConnell and others, 2001). The timescales were further confirmed with beta radioactivity horizons from thermonuclear bomb tests and by identification of known volcanic eruptions (Mosley-Thompson and others, 2001, 2003).

Table 1. Information for all ice cores discussed in the text

Core name	Lat.	Long.	Elevation m	Year drilled	Bottom depth m	Bottom age	Ave. accum. m w.e.	Source
GITS	77.14° N	61.09° W	1887	1996	120.50	AD 1745	0.350	M-T (2001)
NASA-U	73.84° N	49.50° W	2369	1995	151.24	AD 1645	0.332	M-T (2001)
Site D2	71.75° N	46.16° W	2640	1999	132.40	AD 1781	0.450	M-T (2003)
Summit GISP2	72.57° N	38.50° W	3210	1990–93	3053.44	>100 000 BP	0.222	Grootes (1993)
Summit Site T	72.58° N	38.45° W	3200	1989	200.00	AD 1212	0.222	M-T (1993)
Site D3	69.80° N	44.00° W	2560	1999	150.73	AD 1740	0.450	M-T (2003)
Raven	69.50° N	44.50° W	2053	1988	120.96	AD 1717	0.326	M-T (2003)
Franz Josef	80.73° N	63.53° E	509	1997	314.80	AD 1225	0.590	Henderson (2002)
Dunde	38.10° N	96.40° E	5325	1987	139.80	~40 000 years*	0.380	Th (1989)
Guliya	35.28° N	81.48° E	6200	1992	308.60	>500 000 years	0.220	Th (1997)
Dasuopu	28.38° N	85.72° E	7200	1997	167.70	~8000 years	1.000	Th (2000)
Huascarán	9.11° S	77.62° W	6048	1993	166.10	~19 000 years	1.300	Th (1995)
Quelccaya	13.93° S	70.83° W	5670	1983	163.60	AD 470	1.150	Th (1986)
Sajama	18.10° S	68.88° W	6542	1997	132.80	~25 000 years	0.440	Th (1998)
Dyer Plateau	70.34° S	64.87° W	2002	1989–90	235.20	AD 1505	0.440	Th (1994)
Siple Station	75.92° S	84.25° W	1054	1985	303.00	AD 1417	0.461	M-T (1992)
Plateau Remote	84.72° S	43.00° E	3330	1986	200.00	~4000 years	~0.040	M-T (1996)

Note: M-T: Mosley-Thompson first author; Th: Thompson first author.
*Timescale under revision.

To facilitate their comparison, each record has been normalized with respect to its entire length (for cores shorter than one millennium) or with respect to the last millennium for the longer cores. The annual *Z* score is the standardized deviation from its respective mean (annual *Z* score = (annual value − record mean)/record standard deviation). The annual *Z* scores are shown (Fig. 1) as unweighted 10 year (decadal) averages plotted at the midpoint of the decade. The goal here is to examine a global (pole-to-pole) array of ice-core-derived ($\delta^{18}O$) proxy temperature histories to assess 20th-century climate changes within a longer-term perspective. Thus, the averages for the pre-1901 period (1900 to the oldest year in the record or AD 1000) are compared with those from the post-1900 part of the record (1901 to the most recent year in the record). It would be ideal if all the records were of equivalent temporal length; however, the time, expense and logistics required to collect ice cores dictates the depth to which cores are recovered at a given site.

Before comparing the $\delta^{18}O$ records and drawing general conclusions, the advantages and limitations of using $\delta^{18}O$ as a proxy for regional air temperature are reviewed briefly. With respect to ice cores, the $\delta^{18}O$–air-temperature (T_a) relationship has been most extensively investigated in Greenland and Antarctica. In Antarctica, a strong linear relationship between $\delta^{18}O$ and T_a has been demonstrated (Aldaz and Deutsch, 1967; Mosley-Thompson, 1992; Peel, 1992; Jouzel, 1999), with slope (α) values ranging from 0.76 to 0.92‰ °C^{-1}. In Greenland the slope of the linear relationship ranges from 0.62 to 0.67‰ °C^{-1} (Dansgaard and others 1973; Johnsen and others, 1989). More recently, using central Greenland borehole temperatures, Cuffey and others (1995) calibrated the paleothermometer, $\delta^{18}O = \alpha T_a + \beta$. They concluded that $\delta^{18}O$ provides a 'faithful' proxy for long-term average temperature at that site (GISP2); however, they caution that it is inappropriate to use a single set of constants (α and β) to infer past climate changes, as they depend upon factors that change with time. Thus, in keeping with the general practice in ice-core paleothermometry (Jouzel and others, 1997; Jouzel, 1999), we adopt the convention for the polar cores such that more negative $\delta^{18}O$ values reflect cooler air temperatures and less negative $\delta^{18}O$ values reflect warmer air temperatures at the time of condensation.

There has been much less investigation of the $\delta^{18}O$–T_a relationship in non-polar ice cores, in part because records have been limited. Rozanski and others (1992, 1993) reviewed an extensive set of in situ data collected over three decades, mainly from mid-latitude, lower-elevation sites. They suggested that as a first approximation, temperature controls the isotopic composition of precipitation at high and mid-latitudes while the amount of precipitation (the amount effect) controls the isotopic composition in tropical regions (Rozanski and others, 1997). They also concluded that the relationship between longer-term changes in $\delta^{18}O$ and T_a for a given location is more appropriate for paleoclimate investigations (vs short-term applications).

Additional studies have addressed the roles of air temperature and the precipitation 'amount effect' in determining the $\delta^{18}O$ values in precipitation (snowfall) from a regional perspective. Yao and others (1996) conducted the first study of the $\delta^{18}O$–T_a relationship in discrete precipitation events at three sites on the Tibetan Plateau. They reported that averaging $\delta^{18}O$ and air temperatures for discrete precipitation events over longer periods (months to years) removed the synoptic effects and pointed to air temperature as the dominant control on longer-term $\delta^{18}O$ variations at individual sites.

Thompson and others (2000) address the conundrum with Andean precipitation whereby the seasonal $\delta^{18}O$–T_a relationship is opposite that in the polar regions. In the Andes, summer snow has more negative $\delta^{18}O$ values than winter snow, and the seasonal $\delta^{18}O$ range can be large, up to 20‰, although the seasonal temperature range is small. Grootes and others (1989) used a three-step model, employing Rayleigh fractionation in each step, to explain the $\delta^{18}O$

values measured on the Quelccaya ice cap. They concluded that air circulation and air mass stability, rather than temperature, determine the seasonal $\delta^{18}O$ cycle over Quelccaya. They did not address the longer-term relationship between $\delta^{18}O$ and T_a in the tropics that appears consistent with that in the higher latitudes (more negative $\delta^{18}O$ implies colder T_a at condensation). Thompson and others (2000) suggest that seasonal changes in the cycle of deep tropical convection modulate the height of the mean condensation level (hence temperature of condensation). Deep summer convection produces precipitation that is more isotopically depleted than winter precipitation that condenses at lower levels in the atmosphere. Henderson and others (1999) and Vuille and others (2003) have investigated the controls on recent snowfall over Sajama, Quelccaya and Huascarán. Henderson and others (1999) note that the spatial distribution of temperature anomalies in the western tropical Atlantic influences atmospheric circulation at 500 hPa and thereby isotopic fractionation over the Amazon Basin, the primary source of precipitation for Huascarán and Quelccaya. Vuille and others (2003) note that their modeled $\delta^{18}O$ values depend strongly on precipitation amount, but conclude that the $\delta^{18}O$ signal in precipitation is influenced by a combination of mechanisms (precipitation amount, temperature, variability of the moisture source and changes in atmospheric circulation). Bradley and others (2003) report strong linkages between sea surface temperatures across the equatorial Pacific Ocean and $\delta^{18}O$ in ice cores from the tropical Andes as well as the Himalaya (Dasuopu glacier). They, like Henderson and others (1999), report a strong link between $\delta^{18}O$ and El Niño–Southern Oscillation (ENSO) variability, which is not surprising as ENSO integrates all the mechanisms mentioned above. The $\delta^{18}O$–T_a relationship on longer timescales (decades to centuries) is of greater importance here. Thompson and others (2000, 2003) present evidence that on longer timescales, $\delta^{18}O$ variations in Andean precipitation more strongly reflect variations in air temperature than precipitation. Nevertheless, this issue is ripe for further investigation, as the dearth of records and their short duration present challenges to fully addressing the controls on the $\delta^{18}O$–T_a relationship at high-elevation sites in the low latitudes. Here the trends in $\delta^{18}O$, when averaged over decadal and longer timescales, are assumed to reflect primarily trends in air temperature rather than in precipitation.

DISCUSSION

Placing the well-documented 20th-century warming of the Earth's globally averaged temperature within a temporal perspective longer than existing meteorological (Jones and Moberg, 2003) and oceanographic (Levitus and others, 2000) observations (~1860 to the present) requires proxy records such as ice-core-derived $\delta^{18}O$ histories. Figure 1 reveals that the averaged $\delta^{18}O$ for post-1900 precipitation is enriched (to differing degrees) relative to that for pre-1901 precipitation in all cores except those from Siple Station, West Antarctica, and Plateau Remote, East Antarctica. The observed regional trends in temperatures and the associated ice-core-derived $\delta^{18}O$ records are discussed below. If the 20th-century $\delta^{18}O$ trends among the ice cores are broadly consistent with the contemporaneous temperature trends for their respective regions, then the earlier portions (pre-20th-century) of the $\delta^{18}O$ records should also reflect regional temperature trends and provide a viable longer-term perspective for the recent warming.

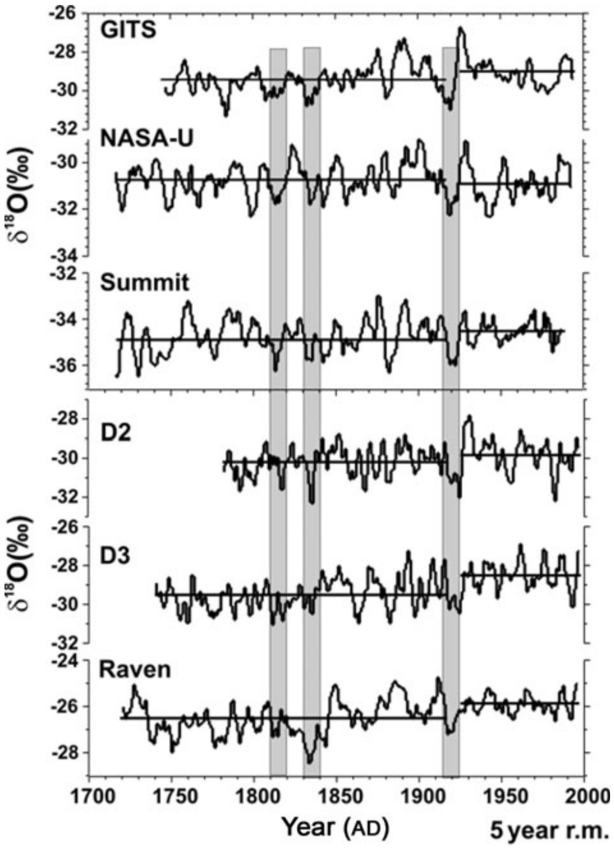

Fig. 2. The 5 year running mean of the annual $\delta^{18}O$ histories from five new Greenland ice cores are shown along with the 1989 Site T record collected at the Summit (GISP2) site. Vertical shaded bars highlight specific time intervals discussed in the text. Horizontal bars in each record show the average $\delta^{18}O$ value before and after the large cooling in 1917–19 that was followed by a rapid, widespread warming over much of the Arctic.

Large regional differences exist among the Greenland $\delta^{18}O$ histories, even between the GISP2 and Summit 1989 records that are ~4 km apart. The 20th-century enrichment appears larger in Summit 1989 than in GISP2, but this does not result from the different record lengths (~500 vs ~1000 years, respectively). To facilitate comparison, the average $\delta^{18}O$ value was calculated for the GISP2 segment that overlaps the Summit 1989 record, and, as Figure 1 reveals, the 500 year $\delta^{18}O$ average is only slightly less than that for the 1000 year segment. Thus, the difference cannot be resolved with data in hand, and likely reflects some combination of differences in dating approaches, sampling schema and real spatial differences among the cores.

The 5 year unweighted running means of the annual $\delta^{18}O$ values (Fig. 2) highlight the regional variability that exists over Greenland on annual to decadal timescales. This variability arises from (1) differences in regional climate forcing, (2) more localized conditions (e.g. an isotopic record is only produced when snow is falling), and (3) post-depositional modification by deflation, drifting and redeposition of surface snow. Despite this regional variability, the coherency of the Greenland $\delta^{18}O$ records (Fig. 2) provides strong evidence that $\delta^{18}O$ of the precipitation (snow) does reflect large-scale, multi-year variations in air temperature.

The vertical bars in Figure 2 highlight three periods of marked isotopic depletion (cooling) that are widely recorded along the west side of the ice sheet from the northwest (GITS) to the southwest (Raven). The two closely spaced shaded bars highlight two decades of well-documented volcanically induced cooling: (1) 1810–20 with the eruptions of Unknown in 1809 and Tambora, Indonesia, in 1815 (Dai and others, 1991; Mosley-Thompson and others, 2003) and (2) 1830–40 with the eruptions of an unidentified volcano in ~1832 and Coseguina, Nicaragua, in 1835 (Cole-Dai and others, 1997). The third bar highlights a brief, but large and spatially coherent, cooling from 1917 to 1919 that is documented in Arctic temperature composites (see e.g. Jones and Moberg, 2003, fig. 2).

This brief cold event preceded an abrupt warming in the high Arctic (60–90° N) from ~1920 to ~1940 (Rogers, 1985). The warming was strongest in the Kara and Barents Sea region and only slightly weaker in western Baffin Bay. It is hypothesized to have been linked to the retreat of sea ice in response to enhanced westerly flow (Bengtsson and others, 2004). Over the ice sheet the post-1920 warming is recorded by slightly enriched $\delta^{18}O$ values at all sites except NASA-U (Fig. 2). Note that the $\delta^{18}O$-inferred warming is most pronounced at the two southernmost sites (Raven and D3). The new Greenland $\delta^{18}O$ records are relatively short and provide a limited temporal perspective, but when coupled with the longer records from Summit they suggest modest 20th-century warming over Greenland with alternating multi-decadal periods of warming and cooling, consistent with the limited available meteorological observations (for a summary see Box, 2002).

An ice-core record from the Windy Dome ice cap in FJL yields quite a different picture (Henderson, 2002). The decadally averaged $\delta^{18}O$ record (Fig. 1) suggests a large warming that began ~1910 and was sustained until mid-century, consistent with observations in the Barents Sea region as discussed above. Longer ice cores from this region are needed to assess the 20th-century warming relative to that associated with the Medieval Warm Period (MWP).

The available high-resolution $\delta^{18}O$ records from Northern Hemisphere high-latitude ice cores are generally consistent with 20th-century observations. Specifically, they record (1) the short-term cooling in response to explosive tropical volcanic eruptions that injected sulfate aerosols into the tropical stratosphere where they were globally distributed, (2) the brief large-scale cooling from 1917 to 1919, and (3) the subsequent rapid (few years), large-scale warming in the Arctic (Rogers, 1985; Overpeck and others, 1997; Jones and Moberg, 2003) that persisted for several decades. Thus it is realistic to assume that $\delta^{18}O$ variations in the pre-20th-century part of the ice-core records also reflect regional temperature trends. The collection of annually dated $\delta^{18}O$ records discussed here suggests that 20th-century warming has been spatially variable over Greenland but, on balance, temperatures have been modestly warmer than in previous centuries, particularly over the southern part of the ice sheet. With the exception of the two southern sites (D3 and Raven), the magnitude of the warming over the 20th century remains smaller than the decadal-scale variability. Eastward in the Russian Arctic, the FJL $\delta^{18}O$ record suggests that 20th-century warming in that high-Arctic region is well outside the norm for the last half-millennium.

Ice cores from tropical and subtropical regions are limited, but new cores are being slowly added to the archive. Two low-latitude ice-core $\delta^{18}O$ composite records have been previously published (Thompson and others, 2003) and are included in Figure 1. The composite for the South American Andes is derived from the Quelccaya, Huascarán and Sajama cores, while the Tibetan Plateau composite is based on the Dunde, Guliya and Dasuopu cores. Note that these two composite $\delta^{18}O$ histories reveal some notable regional differences over most of the last millennium. The Andean composite shows a clear Little Ice Age or recent neoglacial cool period (AD ~1450–1880) as well as warmer conditions from AD ~1100 to 1350, correlative with the MWP. The Tibetan composite shows neither of these multi-century climate variations. The only consistent, century-scale feature between the two regional composites is the marked isotopic enrichment in 20th-century precipitation. Although high-resolution, well-dated tropical and subtropical ice-core records are scarce, they paint a consistent picture, specifically, that the 20th-century warming in the Andes and over the Tibetan Plateau is now outside the range of natural variability for the last millennium. The recent warming in lower-latitude, high-elevation regions is consistent with atmospheric temperatures (Diaz and others, 2003; Jones and Moberg, 2003), Intergovernmental Panel on Climate Change model predictions (Cubasch and others, 2001) and the widespread recession of alpine glaciers (Thompson and others, 2003; Oerlemans, 2005).

The number of annually resolved and carefully dated $\delta^{18}O$ records available from Antarctica is also limited. Three $\delta^{18}O$ records, each representing a different part of the continent, are shown in Figure 1. The annually dated Dyer Plateau (DP) core (Thompson and others, 1994) is located along the spine of the Antarctic Peninsula, a region where temperatures have increased strongly since the 1950s (Marshall and others, 2002; King and others, 2003). Temperature records here are short, with observations extending back to the late 1940s. The decadally averaged 500 year $\delta^{18}O$ history suggests cooler temperatures from AD ~1700 to 1950, followed by ^{18}O enrichment that is contemporaneous with the observed warming in the region (Mosley-Thompson and Thompson, 2003). The 20th-century average $\delta^{18}O$ value exceeds that for the previous 400 years, and the isotopically inferred warming of the last few decades lies well outside the range of natural variability for the last half-millennium.

On the high dry polar plateaus of East and West Antarctica, $\delta^{18}O$ records are available from Plateau Remote and Siple Stations. At Siple Station (SS), the high annual accumulation (461 mm w.e.) results in an excellently preserved and easily interpretable annual record (Mosley-Thompson, 1992). At Plateau Remote (PR), the low accumulation (~40 mm w.e.) makes annual resolution difficult, but well-known time-stratigraphic volcanic horizons provide excellent time constraints for the last millennium (Mosley-Thompson, 1996; Cole-Dai and others, 2000). The $\delta^{18}O$ history at PR shows a very modest ^{18}O depletion (cooling) in 20th-century precipitation relative to that over the last millennium, while the ^{18}O depletion is stronger at SS. Both suggest a recent cooling relative to the earlier part of the record. Antarctic meteorological records are short, few in number and biased toward coastal stations and stations in the Peninsula (Turner and others, 2005). Two long records exist for the interior of the continent (Vostok and Amundsen–Scott South Pole Station) and both show long-term cooling trends. At South Pole (SP), average annual near-surface temperatures have declined slowly by ~0.5°C since record keeping began in 1958 although interannual variability is high (data provided to E.M.-T. by SP meteorological personnel in July 2005).

Thus, recent $\delta^{18}O$ trends in Antarctic precipitation appear regionally consistent with the observed temperature trends. The recent ^{18}O enrichment in the DP core from the spine of the Peninsula and the modest ^{18}O depletion in the SS and PR cores on the high polar plateau are consistent with the warming and cooling trends observed in those regions, respectively (Vaughan and others, 2001 and references therein; Thompson and Solomon, 2002; Turner and others, 2005). Assuming that the earlier portions of these $\delta^{18}O$ records also provide credible temperature proxies, the recent warming in the Peninsula region is unusual within the longer (500 years) perspective available, and the slight cooling in East and West Antarctica lies within the range of past variability (1000 years for PR and 500 years for SS).

CONCLUSIONS

High temporally resolved ice-core-derived $\delta^{18}O$ records have been examined to assess whether they provide reliable proxies for examining regional 20th-century temperature trends within a longer-term perspective. The largest isotopically inferred warming is found in the Russian Arctic. Warmer 20th-century temperatures characterize the Tibetan Plateau and the South American Andes and extend into the Antarctic Peninsula. The 20th-century warming in these regions lies outside the range of natural variability as discerned from the earlier part of their respective $\delta^{18}O$ records. Over Greenland a modest warming that is most pronounced over the southwestern part of the ice sheet has ensued since ~1920. Modest cooling has dominated 20th-century temperatures over the high plateaus of East and West Antarctica. The unique climate histories preserved in the glaciers and ice caps from the Russian Arctic through the tropics to the Antarctic Peninsula may soon be degraded and/or lost if the current warming in those regions persists.

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