Comparison of Lower-Tropospheric Temperature Climatologies and Trends at Low and High Elevation Radiosonde Sites

Dian J. Seidel and Melissa Free

Air Resources Laboratory (R/ARL)

National Oceanic and Atmospheric Administration

1315 East West Highway

Silver Spring, Maryland, USA

dian.seidel@noaa.gov

phone: 301-713-0295 ext. 126

fax: 301-713-0119

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Abstract

Observations of rapid retreat of tropical mountain glaciers over the past two decades seem superficially at odds with observations of little or no warming of the tropical lower troposphere during this period. To better understand the nature of temperature and atmospheric freezing level variability in mountain regions, on seasonal to multidecadal time scales, this paper examines long-term surface and upper-air temperature observations from a global network of 26 pairs of radiosonde stations. Temperature data from high and low elevation stations are compared at four levels: the surface, the elevation of the mountain station surface, 1 km above the mountain station, and 2 km above the mountain station.

Climatological temperature differences between mountain and low elevation sites show diurnal and seasonal structure, as well as latitudinal and elevational differences. Atmospheric freezing-level heights tend to decrease with increasing latitude, although maximum heights are found well north of the equator, over the Tibetan Plateau. Correlations of interannual anomalies of temperature between paired high and low elevation sites are relatively high at 1 or 2 km above the mountain station. But at the elevation of the station, or at the two surface elevations, correlations are lower, indicating decoupling of the boundary layer air from the free troposphere.

Trends in temperature and freezing-level height are generally upward, both during 1979-2000 and during longer periods extending back to the late 1950's. However, some negative trends were found at extratropical locations. In many cases, statistically significant differences were found in trends at paired high and low elevation stations, with tropical pairs revealing more warming (and greater increases in freezing-level height) at mountain stations than at low elevations. This result is consistent with both the observed retreat of tropical glaciers and the minimal change in tropics-wide tropospheric temperatures over the past two decades.

Overall, the analysis suggests that, on diurnal, seasonal, interannual, and multidecadal time scales, temperature variations at mountain locations differ significantly from those at relatively nearby (a few hundred kilometers) low elevation stations. These differences are greatest at the two surface levels, but can persist up to 2 km above the mountain site. Therefore, to determine the nature of climate variability at high elevation sites requires local observations, since large-scale patterns derived from low elevation observations may not be representative of the mountain regions. Conversely, temperature change in mountain regions should not be viewed as necessarily representative of global surface or tropospheric trends.

Introduction

Studies documenting the rapid retreat of mountain glaciers and ice caps around the world during the twentieth century have been cited as proxy evidence of global warming of surface air temperatures (Folland et al., 2001). In the past 20 years, while glaciers in the tropics, in particular, have retreated (Thompson, 2000, Thompson et al., 1993), surface air temperature data show marked upward trends (Folland et al., 2001). However, at mid-tropospheric levels, comparable to the elevations of mountain ice caps and glaciers, satellite and radiosonde temperature data indicate little or no change in tropical temperatures (NRC, 2000). This apparent discrepancy has, to some extent, been reconciled by recognizing that, over short periods of observation, and using imperfect data, one might expect different trends at different levels, especially since the vertical profile of atmospheric temperature change depends on the nature of the climate forcings (NRC, 2000). On large spatial scales, radiosonde observations suggest significantly different trends since 1979 at the surface and free troposphere, especially in the tropics, with associated changes in lower tropospheric lapse rates (Brown et al., 2000, Gaffen et al., 2000). However, it is unclear whether, at mountain sites, temperature variations and trends more closely resemble those in the free troposphere at the same elevation or those at the surface at lower elevation.

Barry's (1992) comprehensive textbook provides excellent reviews of studies, spanning the past century or more, of the atmospheric temperature structure above mountains and plateaus compared with low elevation sites. Many of these studies are based on analysis of surface meteorological data in particular mountain or plateau regions, and it is difficult to draw general conclusions. Although temperature generally decreases with height, lapse rates in the free air do not generally predict altitudinal gradients of surface temperature (Barry, 1992). Temperature differences at a given elevation between the free troposphere and elevated surfaces appear to have diurnal and (particularly in the extratropics) seasonal structure and to depend on meteorological conditions (e.g., cloudiness, winds, lapse rate, humidity, snow cover) and on the profile and slope of the terrain (Barry, 1992). A conceptual model (Tabony, 1985) suggests that the effect of topography on vertical temperature profiles varies from isolated mountain to limited plateau to more extensive plateau.

Several field experiments (Samson, 1965; McCutchan, 1983; Richner and Phillips, 1984; Tabony, 1985 and references therein; see also Barry, 1992) have compared surface temperatures at high elevation (and along slopes) to upper-air temperatures measured by nearby radiosondes. Again, results from different regions are mixed, with some finding mountain temperatures greater than, and others less than, free-air temperatures at the same heights. In some cases, diurnal or, less frequently, seasonal structure to the difference is reported; however, to our knowledge, no long-term, systematic comparisons have been made.

Using a one-dimensional radiative-convective equilibrium model, Molnar and Emanuel (1999) simulated idealized steady-state diurnal-mean, annual-mean temperature and humidity profiles. They obtained notably higher temperatures at the same pressure over elevated surfaces than over sea level, and a decrease of near-surface air temperature with height of the surface of about 2 K/km, much less than the moist or dry adiabatic lapse rates found in free air. Thus simulated upper-tropospheric temperatures over elevated surfaces were substantially higher than over sea-level surfaces.

These observational and theoretical studies address "instantaneous" or climatological-mean differences, not differences in interannual or longer-term variations. However, a number of recent studies have examined trends in temperature at high elevation sites in selected regions. In an overview of climate changes at high elevation sites, Beniston et al. (1997) report 20th century warming of surface temperatures in the Alps exceeding observed global mean warming. Shrestha et al. (1999) report trends in daily maximum temperature at the surface at 49 stations in Nepal, ranging from 72 to 3505 m elevation, including a general cooling (or no trend) during the 1960s, followed by warming from the mid-1970s to mid-1990s, with the greatest warming at the highest stations. Liu and Chen (2000) examined temperature trends at 97 stations located above 2000 m on the Tibetan plateau and report warming during 1955-1996, more pronounced in winter, and greater at the higher elevation stations. Vuille and Bradley (2000) examined temperature data from a network of 268 stations (from 0 to 5000 m) in the tropical Andes region during 1939-1998. Noting an El Niño influence in regional temperature variations, they found general warming trends in annual mean data, with accelerated warming after the mid-1970s, a suggestion of larger trends along the western than the eastern slope, and a decrease in the magnitude of the trend with height, in contrast with the results from Nepal, the Tibetan plateau, and the Alps. Pepin (2000) also finds a decrease in surface temperature trend with altitude during 1952-1997 in the Colorado Front Range, with warming below the treeline and cooling above.

Diaz and Bradley (1997) examined the elevation-dependence of surface temperatures and their trends in ten regions in Europe, Asia, and North and South America. Warming trends in minimum temperatures at high elevation sites during 1951-1989 were equal to or exceeded those nearer sea level. Warming trends in maximum temperatures were smaller than trends in minima, and smaller above 2000 m than at lower elevation. However, these trend differences with elevation were not always statistically significant.

Turning to free-atmospheric trends, and focusing on the tropical belt, Diaz and Graham (1996) examined data from 65 radiosonde stations and found a temporal increase in the free-atmospheric freezing-level height during 1970-1986, and, in a 10-station South American network, during 1958-1990. These freezing level changes were corroborated by surface temperature data from tropical stations at elevations above 1000 m. They were also well simulated by an atmospheric general circulation model, forced by observed sea surface temperatures. Using a somewhat more comprehensive radiosonde station network, Gaffen et al. (2000) noted that, after an abrupt increase in 1976-77 in tropical freezing-level heights, they decreased slightly during 1979-97. During 1979-97, midtropospheric temperatures cooled slightly, while surface temperatures increased significantly, in association with an increase in lower-tropospheric lapse rates. During a longer period, 1960-1997, tropical surface and tropospheric temperature warmed at about the same rate, and freezing levels rose (largely due to the abrupt change mentioned above).

To better understand the apparent discrepancy between recent (past 20 years) glacial retreat and lack of large-scale warming of the tropical lower troposphere, we have used radiosonde observations to investigate the following questions:

1. How are the seasonal, interannual, and multi-decadal temperature variations at the surface at high elevations related to nearby low-elevation surface temperatures?
2. How are temperature variations at the surface at high elevation related to nearby free-tropospheric temperature variations at the same elevation and above?
3. How are trends in atmospheric freezing levels related to temperature trends?

Our focus on temperature notwithstanding, it is important to recognize that, particularly in the dry subtropics, glaciers may be more sensitive to variations in humidity, precipitation, and cloud cover than to temperature change (M. Vuille, personal communication, 2002). Recent work by Wielicki et al. (2002) showing changes in the energy budget of the tropics over the past few decades is consistent with both the observed tropical lapse rate increase (Gaffen et al. 2000) and a more vigorous tropical circulation (Chen et al., 2002), including increased subsidence in the subtropics, which could contribute to glacier retreat by reducing cloudiness and precipitation.

The next section describes the radiosonde data and methods we used in this analysis. Then we compare the climatological features of temperature and freezing levels at high and low elevation radiosonde sites. This is followed by a comparison of interannual variability and trends. A discussion section presents some caveats relevant to this study and is followed by a summary of our main conclusions.

Data and Methods

Using the Comprehensive Aerological Reference Data Set from the NOAA National Climatic Data Center, the most complete archive of global radiosonde data, we have selected pairs of stations with long and reasonably complete records to allow direct comparison between high elevation sites (above 750 m) and low elevation sites within 1000 km. In some cases, high elevation sites are paired with two low elevation sites on different sides of the mountains, with potentially different climatological features.

The resulting network, shown in Figure 1 and Table I, includes 22 pairs of stations that sample some of the world's major continental mountain ranges and high elevation regions, including the Sierra de Guadarrama in Spain, the Anatolian Plateau in Turkey, the Asir Mountains of Saudi Arabia, the Great Escarpment in South Africa, the Yunnan Plateau in China, the Tian Shan Mountains in Kyrgyzstan, the Yablonovyy Khrebet of Siberia, the Rocky Mountains in the United States and Canada, the Sierra Madre in Mexico, the Cordillera de Talamanca in Costa Rica, the Andes in South America, and the highlands of southeastern Brazil. As indicated in Table I, four additional station pairs, with records too short for trend analysis but long enough for climatological comparison, sample the Swiss/German Alps and the Tibetan plateau and allow enhanced sampling of the Tian Shan and Yunnan regions.

Daily soundings were used to create 00 and 12 UTC monthly means of temperature at several levels in the lower troposphere, and of the height of the freezing level. For the 22 station pairs used to analyze trends, data for a given month were used only if available for both stations in a given pair, so that climatological mean values and monthly anomaly time series are based on the same period of observation. For the four station pairs at which only climatological comparisons were made, we required at least ten years of data from each station, but not necessarily the same years. The temporal homogeneity of the station time series was checked visually, and some records were shortened to avoid sudden jumps in temperature associated with changes in station elevation.

As shown schematically in Figure 2a, we make five sets of comparisons: surface temperature at low and high elevations; temperature at the elevation of the mountain site and at the same elevation above the low site; temperatures at 1 and 2 km above the mountain site and at the same elevation above the low site; and heights of the freezing level at both sites. With the exception of the surface data (and, in some cases the freezing level), values were obtained by interpolating between reported levels. Such interpolation is justified by the radiosonde data reporting procedures which specify that a so-called significant level be reported if there is substantial departure from linearity between two mandatory reporting levels.

As suggested in Figure 2, and the elevation data in Table I, radiosonde stations in mountain regions are generally not located on mountain peaks. Rather, they are near cities or at airports on plateaus or in valley regions, so the data may be influenced by small-scale features and imperfectly represent mountaintop conditions at which glaciers might reside.

Climatological Features

The following examination of the climatological features of temperatures and freezing levels at the stations in the network begins with a simple example in which the mountain site is generally cooler than the low elevation site. We then demonstrate that this simple pattern is not common. First we discuss diurnal patterns to the temperature differences and then demonstrate that seasonal patterns prevail at many locations and give examples. The section ends with a discussion of the climatology of the freezing level.

Conventional wisdom holds, and some observational studies (Samson, 1965; Pepin, 2000) would confirm, that temperatures at high elevation are lower than at low elevation, particularly in summer, when mountains are viewed as retreats from the heat. Buttressing Barry's (1992) analysis of the literature, our comparison of climatological annual temperature cycles at the pairs of radiosonde stations shown in Figure 1 only partially supports this notion. Figure 3 shows the annual cycles of temperature, and temperature differences, at three sample station pairs that typify patterns at other locations in the network. For each pair, we show climatological 00 UTC temperatures at the low and high elevation station, at each of the levels shown in Figure 2a, and the 00 and 12 UTC differences. For each pair, the approximate local standard time of the observations is given. The salient characteristics of these pairs are summarized in Figure 2b-d.

The first pair (top row of Figure 3) is from the "Brazil 1" region (Table I) represented by Rio de Janeiro at 42 m elevation and Curitiba at 908 m. As expected, temperature decreases with elevation, and is higher in summer (December-January-February, DJF), at both stations. The temperature difference at the surface (open circles) shows Curitiba to be 6 to 8 K cooler all year long, and that difference is consistent at 00 and 12 UTC (evening and morning). Curitiba surface temperatures are also a few (1 to 5) degrees lower than temperatures over Rio de Janeiro at the altitude of the Curitiba surface data, 908 m (closed circles), with larger differences in winter. The sign of the temperature difference remains negative at 1 km and 2 km above the Curitiba station, but the magnitude is significantly diminished to about 1 K. The freezing level at Rio de Janeiro resides between the 4 km and 5 km level (Figure 4b), and it is about 100-150 meters lower over Curitiba, consistent with the lower temperatures there.

Within our network, only a few station pairs share the attributes of "Brazil 1" station pair. These attributes, summarized in Figure 2b are: equal or lower temperatures over the high elevation site all year long, particularly at the surface and mountain levels; lower freezing-level heights over the high elevation site; similar temperature differences at 00 and 12 UTC; and similar temperature differences in each calendar month. The station pairs in this group include Brazil 1, Andes 2, Brazil 2, Madre 2 (for which only 12 UTC data were available) and Talamanca.

Figure 4 summarizes the annual-mean temperature differences and climatological freezing heights, and their differences, for all the station pairs, arranged by latitude (Figure 4d). The five station pairs listed above all fall within the tropical zone (within 30 degrees latitude of the equator), but they are not the only pairs in that zone.

A second set of five other tropical station pairs (Asir, Escarpment, Madre 1, Tibet, and Yunnan 1) have the distinctive feature that the mountain location is warmer in the afternoon than the air above the low elevation site (Figure 2c). For this second tropical group, the difference in elevation between the two stations averages 1891 m, compared to an average difference of 1006 m in the first tropical group.

This feature is also evident at many extratropical station pairs, such as "Anatolia 2" (middle row of Figure 3) where climatological temperature differences differ markedly from day to night. At night (00 UTC), surface temperatures at Ankara (894 m) are 3 to 6 K lower than at Istanbul (40 m), and at the 894 m level (the surface level at Ankara) and above, the differences are much smaller. During the daytime (12 UTC), however, Ankara surface temperatures are up to 10 K higher than at the same altitude over Istanbul. The warmer air persists at 1 km above the Ankara surface, but by 2 km above, the differences diminish to nighttime levels. Although, on annual average, surface temperatures are lower at Ankara at both observation times, they are higher at the 1 km and 2 km level (Figure 4a), which leads to somewhat higher freezing levels at Ankara than Istanbul (Figure 4c). Similar diurnal differences in temperature, with high elevation sites warmer in daytime and colder at night than above the low elevation site, are found at the Alps, Anatolia 1, Anatolia 2, Andes 1, Cascades, Guadarrama, N. China, all four Rockies regions, Tian Shan 1 and 2, and Yunnan 1 and 2, as well as the five tropical pairs previously mentioned. Four other station pairs did not have sufficient observations at both 00 and 12 UTC to make a comparison. Thus the majority of the extratropical regions, and those tropical station pairs with relatively large differences in elevation, exhibit this marked day-night difference (Figure 2c).

To further explore the diurnal variability of the temperature differences, we are limited by the twice-daily radiosonde observing schedule. However, the global network of station pairs span a wide longitudinal range (Figure 1), and so the local times of observation vary across the diurnal cycle. Figure 5 shows the annual average temperature difference at each level for all the station pairs as a function of local time of observation, with many pairs appearing twice because both 00 and 12 UTC data were available (Table I). As in Figure 4a, Figure 5 indicates that the temperature differences are smaller at the 1 km and 2 km levels than below, and that surface temperature differences are generally (but not always) negative, with the mountain site cooler than the low elevation site. In Figure 5, at these three levels (surface, 1 km and 2 km) there is no obvious relationship between temperature difference and local time of observation. But at the elevation of the mountain, the data (filled circles and sine wave fit) suggest a diurnal signal, with the mountain site warmer from noon until evening, and the low elevation site warmer from after midnight until late morning.

This result is not fully consistent with McCutchan (1983), who compared radiosonde data at 04, 10 and 16 LST to surface data from the San Bernadino Mountains in southern California during June to October 1975. Surface temperatures were lower than free-air values at 04 LST, but higher at 10 (in contrast to our results) and at 16 LST. Using data from eight mountaintop observatories and five radiosonde stations (making observations up to five times per day) in the Alps for May to August 1982, Richner and Phillips (1984) find the mountain surface temperature to be higher than the free-air temperature from about 10 LST to 19 LST and lower during the rest of the day. The daytime differences average about 2K and were generally greater than at night (Richner and Phillips, 1984). Similarly, Samson (1965) compared temperatures at Pikes Peak to radiosonde data from Denver, Colorado, during July and August 1963, and found consistently warmer mountaintop temperatures at 11 LST and cooler mountaintop temperatures at 23 LST, with larger differences in daytime. Our results (Figure 5) also indicate larger differences during daytime than nighttime, but of larger magnitude than either Samson (1965) or Richner and Phillips (1984).

A third common feature, notable for many extratropical station pairs, is a marked annual cycle in the temperature differences as exemplified in Figure 3 (bottom row) by the “Rockies 1” pair. Temperatures at the low elevation station, Norman Wells (64 m), show a pronounced annual cycle and climatological inversions from the surface to 1 km above the high elevation station, Whitehorse (704 m), in winter. Both the inversion strength and the amplitude of the annual cycle are reduced at Whitehorse. Consequently, the temperature differences have a more complex seasonal variation than at the Brazil 1 pair (top row of Figure 3). In general, temperatures above Whitehorse exceed those above Norman Wells, which might be expected as Norman Wells is about 5 degrees latitude closer to the Arctic Circle. The largest differences in spring and fall occur at the surface and at the elevation of the surface at Whitehorse. In summer, however, Whitehorse is cooler at the surface both during the day (00 UTC, 15 LST) and at night (12 UTC, 03 LST) than the surface at Norman Wells. Whitehorse surface temperatures are higher than Norman Wells upper-air temperatures at the elevation of Whitehorse during daytime in all months, and at night in all months except summer (Figure 2d). Freezing levels at Norman Wells vary between about 2.5 km in summer to a few hundred meters in winter (Figure 4b), and they are about 0.5 km higher at Whitehorse. The marked seasonal variation of the temperature differences is also seen in the following regions: Cascades, Guadarrama, Rockies 2 and 3, Tian Shan 2, and Yunnan 2.

The main climatological features of the height of the freezing level are shown in Figure 4, where seasonal mean values at the low elevation sites are plotted, in most cases for both 00 and 12 UTC[1]. At these low elevation sites, the freezing level varies in height from a few hundred meters in winter at Beijing (N. China), Norman Wells (Rockies 1), and Kirensk (Yablonovyy) to more than 5 km in summer over Guilin (Yunnan 1) and Xichang (Tibet). At the high elevation sites (not shown) the freezing levels are often higher (Figure 4c), with summertime average values of 6043 m at Lhasa (Tibet). During the equinoctial seasons (MAM and SON), freezing levels are near 4.5 km in the broad tropical belt from 30N to 30S, but the highest values are seen at the northern limit of this range, over Tibet in summer, in association with evening (20 LST) surface temperatures of 20 deg. C (compared with 11 deg C at 08 LST) at Lhasa. Mexico City, at lower latitude (19N compared with 29N for Lhasa), also has average JJA surface temperatures of 20 deg. C (at 18 LST), but the freezing level at Mexico City in JJA is at 4760 m, about 1.3 km lower than at Lhasa. This discrepancy is likely due to the larger spatial extent of the Tibetan Plateau compared with the Sierra Madre, and with all the other mountain regions in this study.

The high freezing level over Tibet is not associated with low free-air lapse rates, as would be suggested by the analysis presented by Tabony (1985). We find lapse rates in the first kilometer over

[1]Freezing levels were taken as the surface whenever the surface temperature and all free-air temperatures were #0EC. At a few high latitude locations, seasonal mean freezing-level heights, therefore, can exceed the surface elevation even when surface temperatures are, on average, below freezing.

Lhasa averaging 6 K/km at 08 LST and 10 K/km at 20 LST, in contrast to Tabony's estimate of 3-6 K/km, based on surface (not upper-air) temperatures at Lhasa and Patna. Thus the high freezing level over Lhasa is due to the high surface temperatures at this extremely high elevation site.

There is a marked dependence of freezing-level height on latitude, which is much more pronounced than in a similar plot (not shown) of temperatures, which have a less consistent latitudinal signal, mainly because of the variability of the station elevations and the fact that our temperature data are determined relative to the mountain station elevation. The seasonal variations in freezing levels are about 2 km in the extratropics, but less than 1 km in the tropical region. There is little difference between the 00 and 12 UTC heights. Similarly, there is little change from 00 to 12 UTC in the difference in heights between the low and high elevation stations.

Our freezing-level height values compare well with those of Harris et al. (2002), who used twenty years of output from the National Center for Environmental Prediction's reanalysis to plot global freezing-level height. Their climatological mean maps show strong zonal symmetry, particularly in the tropics, where freezing-level heights are generally between 4.5 and 5.0 km. As in our data, July values are highest over the Tibetan Plateau, where they exceed 6 km.

Trends

Having established that the diurnal and seasonal variations of temperature and freezing level can be notably different above high and low elevation locations in a given mountain region, we now turn our attention to a comparison of interannual variations and multidecadal trends in these regions. To ascertain association between the interannual variability at the high and low elevation sites, we computed linear correlation coefficients of detrended temperature monthly anomalies at each of the levels shown in Figure

2. Figure 6 shows these correlations as a function of station separation. At the surface and the mountain elevation, correlations can be rather low, although typical values are about 0.5. The data are more closely correlated at the 1 km and 2 km level, where values are generally between 0.6 and 0.9. As expected, correlations at these higher levels tend to decrease with increasing station separation.

However, not all station pairs within a few hundred kilometers demonstrate high correlations at these higher levels. In fact, some potential station pairs were considered but not included in this study because of low correlation ($r < 0.5$) of monthly temperature anomalies even at the 2 km level. These include Chilean and Argentine stations in the southern Andes, the northern Andes region represented by Bogotá and San Andres Island, stations in the Atlas Mountains of Algeria, and stations in the Himalaya and in southern India.

Temperature and freezing level trends were computed for most of the remaining station pairs, for two different data periods. The first period represents the longest period of reasonably complete data for a given station pair, and is given in Table I. These periods vary from region to region, but generally cover the late 1960's through 2000 and were defined as the longest period, exceeding 20 years, for which at least 60% of the months had data available. For each observation time, we required at least ten observations per month, although the number of missing days was generally quite low, as would be expected for these lower tropospheric observations. The second data period for trends was restricted to the 1979-2000 period, which coincides with the MSU satellite data period and with the period of rapid tropical glacier retreat.

Figure 7 (a-d) shows temperature trends for the longer periods, as a function of latitude, for the mountain and low elevation sites. In some cases, two sets of trends are shown, if both observation times were used (Table I). We consider statistically significant those trends whose 95% confidence intervals do not span 0, and whose p value is < 0.05. Based on these criteria, the surface trends (Figure 7a), most of which indicate warming, are generally not statistically significant. Seven of the 32 low station surface trends, and eleven of the mountain surface trends, are statistically significant, and in all but one case (Denver at 12 UTC), they are positive. The largest trends, both positive and negative, tend to be at the mountain stations. The most striking surface warming (exceeding 0.3 K/decade) is at the mountain stations of Mexico City (Madre 1 and 2), Mendoza Airport (Andes 1), Curitiba (Brazil 1), Sao Paulo (Brazil 2), Madrid (Guadarrama), San Jose (Talamanca), Frunze Bishkek (Tian Shan 1), Lhasa (Tibet), and Whitehorse (Rockies 1).

To better assess the differences between the low and high elevation station trends, we computed the trend in the difference (high station minus low station) monthly anomaly time series. Compared with examining the trend difference, this approach is more likely to yield statistically significant results, since the strong covariability (Figure 6) of the two time series is removed (Santer et al., 2000). The difference trends, for each of the four levels, are shown in Figure 7e. Thirteen of the 32 surface difference trends (filled circles) are statistically significant. These indicate a tendency for greater warming at the mountain station (positive trend) in the tropics and a mixed pattern in the extratropics. Similar features are seen in the comparison of trends at the elevation of the mountain station (Figures 7b and 7e).

At the 1 and 2 km levels, however, there is less discrepancy between the high and low elevation station trends than seen at lower levels. For the most part, the trends at these levels are near-zero or upward (Figures 7c and 7d), and about half the trends shown are statistically significant. There remains a tendency for the tropical mountain sites to show greater warming than the low elevation sites, with the opposite pattern at Ankara (Anatolia 2), Krasnyy Chikoy (Yablonovyy), Bloemfontein (Escarpment) and Salta (Andes 2).

Figure 7f shows temperature trends at the elevation of the mountain station for the shorter data period 1979-2000. For this more recent period, the warming trends in the northern extratropics are larger than for the longer period (Figure 7b). Statistically significant difference trends (not shown) indicate more positive trends at the mountain sites (by about 0.5 to 0.8 K/decade) in the Madre 1, Madre 2, and Brazil 2 regions, and lower trends at the mountain sites (by 0.7 to 1.8 K/decade) in the Guadarrama, Anatolia 2, Yablonovyy, and Rockies 3 regions.

How are freezing level variations and trends related to temperature? Correlations of time series of monthly mean freezing levels with monthly mean temperatures at each level are generally very high. The correlation with surface temperature exceeds 0.85 in 80% of the cases, and the correlation with temperature at higher elevation exceeds 0.85 in more than 90% of the cases. More than three-fourths of the correlations with temperature above the surface exceed 0.95. The lower correlations are at the tropical stations, where the lack of a prominent annual cycle in temperature and freezing levels reduce the correlation.

Just as freezing level variations are closely tied to (upper-air more than surface) temperature variations, so are freezing level trends well correlated with upper-air temperature trends. Figure 8 shows scatterplots of temperature and freezing level trends for the longer data periods. The top panel involves temperature trends at the 2 km level, and the lower panel involves surface temperature trends. Clearly, the latter are much poorer predictors of freezing level trends than the former. The data in Figure 8 suggest that a 1 K/decade warming at the 2 km level is associated with an approximate 125 m/decade rise in freezing level. On average in the tropics (based on 58 radiosonde stations in the 30N-30S latitude zone), Gaffen et al. (2000) found an approximate 30 m/decade rise in freezing level during 1960-1997, in association with an approximate 0.2 K/decade warming at 700 hPa, which is reasonably consistent with the data in Figure

8.

The pattern of freezing level trends (not shown) for the longer trend periods (of varying lengths, given in Table 1) resembles the pattern of temperature trends at 1 and 2 km (Figures 7c and 7d). We find near-zero trends at tropical low elevation stations, and freezing-level height increases of approximately 20 to 70 m/decade at tropical high elevation stations. The largest increases in freezing-level height (50 to 100 m/decade) were at Madrid (Guadarrama), Salta (Andes 3), Mexico City (Madre 1 and 2), Quillayute (Cascades), and Topeka (Rockies 3). Comparably large decreases were found at Samsun (Anatolia 1), Durban (Escarpment), and Buenos Aires (Andes 1). Of the 31 pairs of trends calculated (from 22 stations, using either daily or twice-daily observations, see Table I), ten reveal statistically significant upward trends in the mountain minus low elevation freezing-level height difference (i.e., greater increase in freezing-level height over the mountain station), and five indicate statistically significant downward trends.

Discussion

The differences in surface and lower-tropospheric temperature climatology and trends between high and low elevation sites highlight the difficulty of assessing global or regional temperature changes using a sparse network, such as the radiosonde network. This study focuses on the impact of topography on the spatial heterogeneity of temperature variations. However, caution is advised in interpreting the differences we find using the radiosonde network. As noted above, and as suggested by Figure 2, our "high" elevation stations are often at relatively low elevation, rarely on mountain peaks (Table I). This poses particular problems in tropical regions, where radiosonde stations may be a few kilometers below glacier equilibrium line altitudes. In the cases of high plateau regions (e.g., Tibet, the Great Escarpment, the Brazilian highlands, the Yunnan Plateau), the radiosonde stations are likely quite representative of the larger surrounding high elevation region. But some of the upland sites are actually in topographic hollows (e.g. Mexico City, Whitehorse, Salt Lake City) in major mountain ranges, and the planetary boundary layer in such complex terrain is likely very spatially heterogeneous. Features such as the development of temperature inversions and cloud cover and the influence of topographically-generated winds, which are highly dependent on synoptic weather conditions, may exhibit different structure and frequency at our sites than at mountain peaks. For example, Pepin and Losleben (2002) find that temperature trends during the second half of the 20th century in the Front Range are quite different in Denver, where the lower troposphere has warmed, and nearby Niwot Ridge, where low elevation warming and high elevation cooling contribute to an increase in lapse rate.

A second feature of our station network is the rather natural tendency for the high elevation sites to be more "continental" than the paired low elevation sites, which are often near coasts. The Guadarrama, Madre 1, Cascades, and Escarpment pairs are notable in this regard. In these cases, differences associated with continentality and elevation are intertwined. Similarly, many of our stations are in urban regions, a natural result of the siting of upper-air stations at airports. The well-known urban heat island may influence climatological temperatures at the sites, particularly at night and in winter. Furthermore, growth in the urban heat island over time could lead to temperature trends that are unrepresentative both of surrounding rural areas and of areas of equivalent elevation. This issue is likely not important much above the boundary layer, so that temperatures 1 and 2 km above the mountain site, and freezing levels, may be less affected than surface values.

Third, our ability to directly address the issue of tropical glacial retreat is compromised by the lack of station pairs in the vicinity of some of the most well-documented glaciers: those on Mts. Kenya and Kilimanjaro in east Africa and those in the tropical Andes.

Lastly, the sparsity of the radiosonde network, combined with the great size of the world's mountain ranges, means that many of our station pairs are hundreds of kilometers apart (Table I). Changes in synoptic weather patterns could contribute to differences in temperature trends that, again, are not related to elevational differences.

Conclusions

Based on comparison of radiosonde temperature and freezing level data at 26 pairs of stations, we find the following patterns.

- Climatological temperature differences between mountain and low elevation sites show latitudinal and elevational differences. At a few station pairs in the tropics where the elevation difference between the high and low sites is relatively small, temperatures over the high site were lower at all levels (from the surface to 2 km above the high site) all year round and both day and night.
- At tropical station pairs with large elevation differences, and at most extratropical station pairs, there is a marked difference from day to night in the pattern of temperature differences, particularly at the elevation of the mountain site. Mountain site temperatures are higher than low elevation site temperatures at the same elevation during daytime, but lower at night.
- At most extratropical locations there is a marked seasonal cycle to the temperature differences.
- Atmospheric freezing-level heights tend to decrease with increasing latitude, although maximum heights are found well north of the equator, over the Tibetan Plateau. Freezing-level heights vary from a few hundred meters at high latitudes in winter to maximum values over 6 km above Tibet in summer, with typical values of 4.5 km during spring and fall in the tropical belt.
- Correlations of interannual anomalies of temperature between paired high and low elevation sites

are relatively high at 1 or 2 km above the mountain station. But at the elevation of the station, or at the two surface elevations, correlations are lower, indicating decoupling of the boundary layer air from the free troposphere.

- Trends in temperature and freezing-level height are generally upward, both during 1979-2000 and during longer periods extending back to the late 1950's. However, some negative trends were found at extratropical locations. In many cases, statistically significant differences were found in trends at paired high and low elevation stations, with tropical pairs revealing more warming at mountain stations than at low elevations.

Overall, this analysis suggests that, on diurnal, seasonal, interannual, and multidecadal time scales, temperature variations at mountain locations can be significantly different from those at relatively nearby (a few hundred kilometers) low elevation stations. These differences are greatest at the two surface levels, but can persist up to 2 km above the mountain site. Therefore, to determine the nature of climate variability at high elevation sites requires local observations, since large-scale patterns derived from low elevation observations may not be representative of the mountain regions.

Nevertheless, our analysis reveals indication of greater warming (and greater increases in freezing-level height) at tropical mountain locations than at similar altitudes above low elevation stations. This result, with all the above-mentioned caveats, provides some explanation for the apparent discrepancy between minimal temperature change in the tropical lower troposphere since 1979 (NRC, 2000; Gaffen et al., 2000) and the observed tropical glacial retreat (Folland et al., 2000; Thompson, 2000).

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Figure Captions

Figure 1. Map of the locations of the station pairs used in this study. See Table 1 for detailed station information for each region.

Figure 2. (a) Schematic drawing of the levels at which temperature (and freezing level) comparisons between high and low elevation radiosonde stations are made in this study. Summaries of the climatological temperature and freezing level differences for three groups of station pairs discussed in the text: (b) a group of five tropical pairs, typified by the "Brazil 1" pair, (c) a group of five different tropical pairs, with larger elevation differences than in the first group, plus many of the extratropical pairs, typified by the "Anatolia 2" pair, and (d) a group of seven extratropical pairs, typified by the "Rockies 1" pair.

Figure 3. Climatological 00 UTC annual temperature cycles (left side) and temperature differences for both 00 and 12 UTC (right side) for three regions: Brazil 1 (top row), Anatolia 2 (middle row), and Rockies 1 (bottom row). For each low elevation station (first panel on left), temperatures are shown for the surface, the elevation of the high site, 1 km above the high elevation, and 2 km above the high elevation. For the high elevation station (second panel on left), temperatures are shown at the surface and at 1 and 2 km above. Temperature differences are for the level pairs shown in Figure 2a.

Figure 4. a) Climatological annual-mean temperature differences (high elevation station minus low elevation station) for four combinations of levels (see Figure 2a) for each region. The regions are arranged in order of increasing latitude, as shown in d). In many cases, two values are given for each region. The value plotted opposite the tick mark is for 00 UTC; the 12 UTC difference is immediately to the right, between tick marks. b) Climatological seasonal-mean freezing levels at the low elevation sites. c) Climatological annual-mean difference (high elevation station minus low elevation station) in the height of the freezing level.

Figure 5. Climatological annual-mean temperature differences at four levels plotted as a function of the local time of observation. The curve is a sine wave fit (with $r^2 = 0.52$) to the data points from the mountain site elevation.

Figure 6. For each station pair, linear correlation of high and low elevation station detrended monthly temperature anomaly time series at four levels, as a function of distance separating the stations.

Figure 7. For each region, temperature trends at the high and low elevation sites, plotted at the average latitude of the two sites, at a) the surface, b) the altitude of the high elevation surface, c) 1 km above the high elevation surface, and d) 2 km above the high elevation surface. The trends in a-d are for the long periods shown in Table I. e) Trends in the temperature differences between the high and low elevation sites for the long periods. f) Trends at the elevation of the high station for the period 1979-2000.

Figure 8. Scatterplot of freezing level trends and temperature trends, for the periods listed in Table 1, at the surface (bottom) and at the level 2 km above the high elevation station. Each point represents one station and one observation time.

Table I. Network of pairs of radiosonde stations used in this study. Each pair is named according to its mountain region. Low and high elevation stations are listed by WMO identification number, name, latitude, longitude and elevation. The difference in the stations' elevations and their separation distance is given. The data used for trend analysis are specified by the observation time (00 or 12 UTC, or both), whether trends were computed both for the (short) satellite data period (1979-2000), a (long) period beginning before 1979, both of these, or neither. The period for trends given is for the long period.

Region	Low Station ID	Low Station	Low Station Elev.	High Station ID	High Station	High Station Elev.	Ob. Time for Trends	Trend Analysis	Period for Trends	Elev. Diff. (m)	Separation (km)
Alps	10338	Hannover (52N, 10E)	57	10921	Neuhausen (48N, 9E)	807	00	none	none	750	500
Anatolia 1	17030	Samsun (41N, 36E)	4	17130	Ankara (40N, 33E)	894	both	both	1966-2000	890	327
Anatolia 2	17062	Istanbul (41N, 30E)	40	17130	Ankara (40N, 33E)	894	both	both	1966-2000	854	340
Andes 1	87576	Buenos Aires (35S, 59W)	20	87418	Mendoza Airport (33S, 69W)	704	12	both	1974-2000	684	971
Andes 2	85442	Antofagasta (23S, 70W)	137	87047	Salta (25S, 65W)	1221	12	both	1973-1999	1084	530
Andes 3	87155	Resistencia (27S, 59W)	52	87047	Salta (25S, 65W)	1221	12	both	1973-1999	1169	704

Region	Low Station ID	Low Station	Low Station Elev.	High Station ID	High Station	High Station Elev.	Ob. Time for Trends	Trend Analysis	Period for Trends	Elev. Diff. (m)	Separation (km)
Asir	40477	Jeddah (22N, 39E)	17	40569	Khamis Mushait (18N, 43E)	2057	both	short	1984-1998	2040	517
Brazil 1	83746	Rio de Janeiro (23S, 43W)	42	83840	Curitiba (26S, 49W)	908	12	both	1965-2000	866	671
Brazil 2	83746	Rio de Janeiro (23S, 43W)	42	83780	Sao Paulo (24S, 47W)	802	12	both	1970-2000	760	359
Cascades	72797	Quillayute (48N, 125W)	56	72681	Boise (44N, 116W)	876	both	both	1966-2000	820	808
Escarpment	68588	Durban (30S, 31E)	8	68442	Bloemfontein (29S, 26E)	1359	both	both	1975-1997	1351	459
Guadarrama	08001	La Coruna (43N, 8W)	67	8221	Madrid (40N, 4W)	638	both	both	1959-2000	571	512
Madre 1	76692	Veracruz (19N, 96W)	12	76679	Mexico City (19N, 99W)	2234	both	both	1968-1998	2222	312
Madre 2	76612	Guadalajara (21N, 103W)	1551	76679	Mexico City (19N, 99W)	2234	12	both	1979-2000	683	470

Region	Low Station ID	Low Station	Low Station Elev.	High Station ID	High Station	High Station Elev.	Ob. Time for Trends	Trend Analysis	Period for Trends	Elev. Diff. (m)	Separation (km)
N. China	54511	Beijing (40N, 116E)	54	53463	Hohhot (40N, 111E)	1062	both	short	1982-2000	1008	421
Rockies 1	71043	Norman Wells (65N, 127W)	95	71964	Whitehorse (61N, 135W)	704	both	both	1963-1983	609	656
Rockies 2	72261	Del Rio (29N, 101W)	333	72365	Albuquerque (35N, 107W)	1620	both	both	1957-2000	1287	834
Rockies 3	72456	Topeka (39N, 96W)	329	72469	Denver (40N, 105W)	1611	both	both	1953-2000	1282	797
Rockies 4	72493	Oakland (38N, 122W)	8	72572	Salt Lake City (41N, 112W)	1288	both	both	1956-2000	1280	943
Talamanca	80001	San Andres Island (13N, 82W)	2	78762	San Jose (10N, 84W)	920	12	long	1972-2000	918	396
Tian Shan 1	38062	Kzyl-Orda (45N, 66E)	131	38353	Frunze Bishkek (43N, 75E)	756	both	long	1960-1992	625	761
Tian Shan 2	38353	Frunze/Bishkek (43N, 75E)	756	36974	Naryn (41N, 76E)	2049	none	none	none	1293	194
Tibet	56571	Xichang (28N, 102E)	1517	55591	Lhasa (29N, 91E)	3649	none	none	none	2132	1110

Region	Low Station ID	Low Station	Low Station Elev.	High Station ID	High Station	High Station Elev.	Ob. Time for Trends	Trend Analysis	Period for Trends	Elev. Diff. (m)	Separation (km)
Yablonovyy	30230	Kirensk (58N, 108E)	261	30935	Krasnyy Chikoy (50N, 109E)	770	both	both	1957-2000	509	824
Yunnan 1	57957	Guilin (25N, 110E)	178	56778	Kunming (25N, 102E)	1891	00	short	1979-1993	1713	772
Yunnan 2	57083	Zhengzhou (34N, 113E)	109	52889	Lanchou (36N, 104E)	1519	none	none	none	1410	904

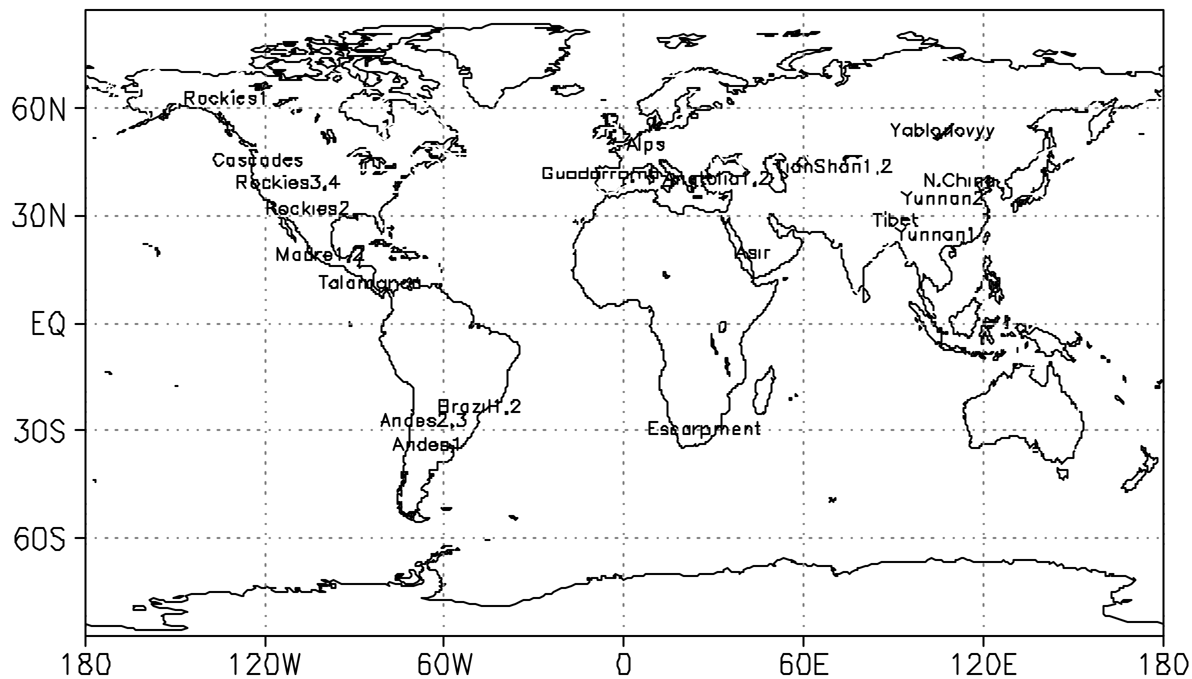

60N
30N
EQ
30S
60S
180
120W
60W
0
60E
120E
180
Rockies1
Cascades
Rockies3,4
Rockies2
Madre1,2
Talamanca
Brazil1,2
Andes2,3
Andes1
Alps
Guadarrama
Anatolia1,2
TianShan1,2
Asir
Escarpment
Yablonovyy
N.China
Yunnan2
Tibet
Yunnan1

Figure 1

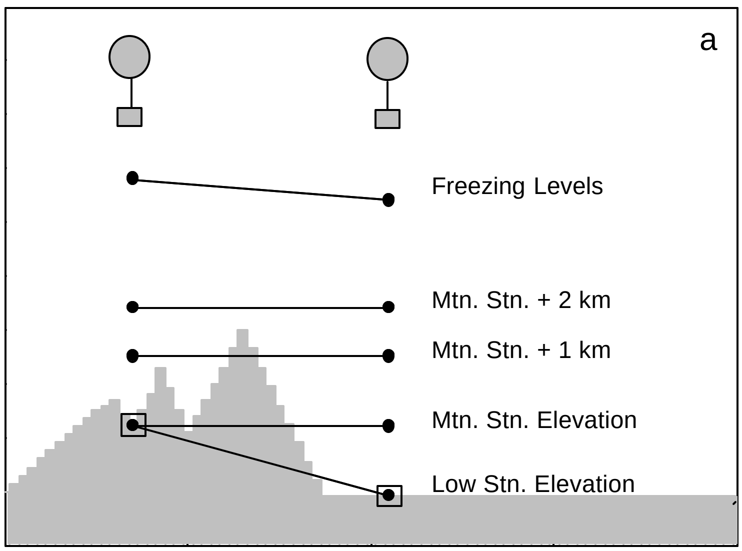

a
Freezing Levels
Mtn. Stn. + 2 km
Mtn. Stn. + 1 km
Mtn. Stn. Elevation
Low Stn. Elevation

Figure 2a

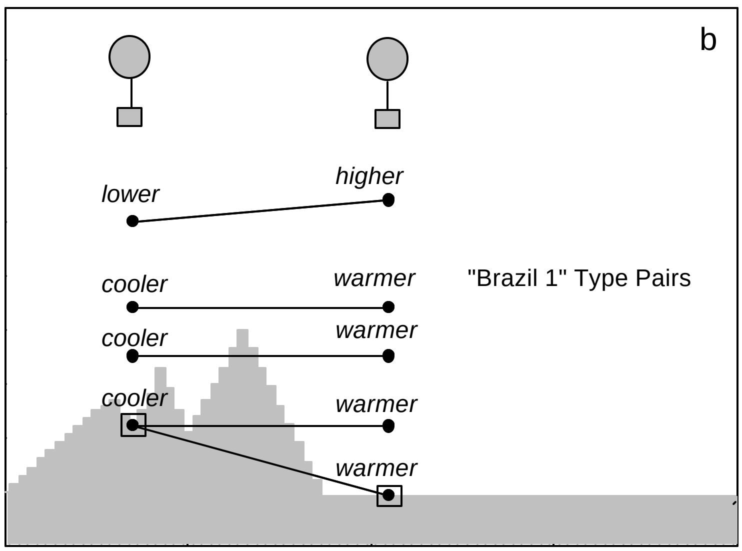

b
lower
higher
cooler
warmer
"Brazil 1" Type Pairs
cooler
warmer
cooler
warmer
warmer

Figure 2b

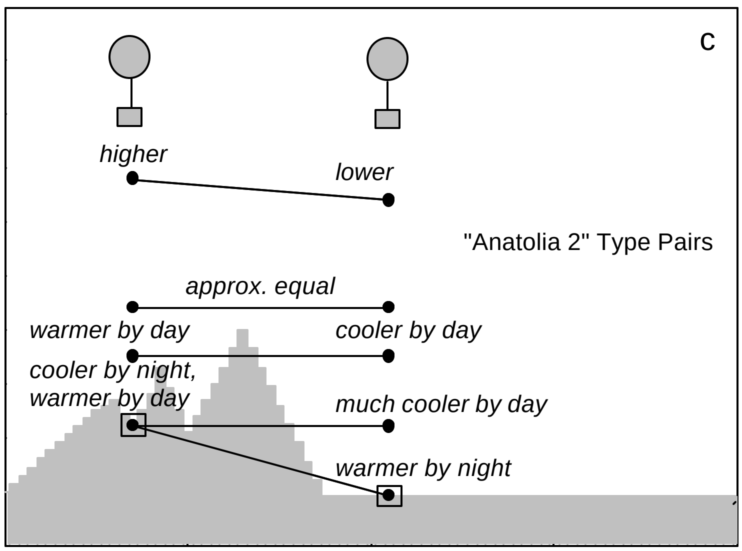

c
higher
lower
"Anatolia 2" Type Pairs
approx. equal
warmer by day
cooler by day
cooler by night,
warmer by day
much cooler by day
warmer by night

Figure 2c

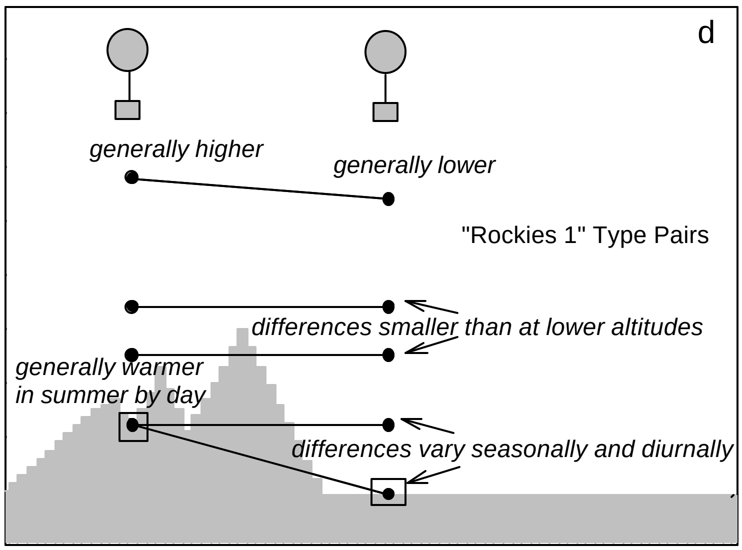

d
generally higher
generally lower
"Rockies 1" Type Pairs
differences smaller than at lower altitudes
generally warmer
in summer by day
differences vary seasonally and diurnally

Figure 2d

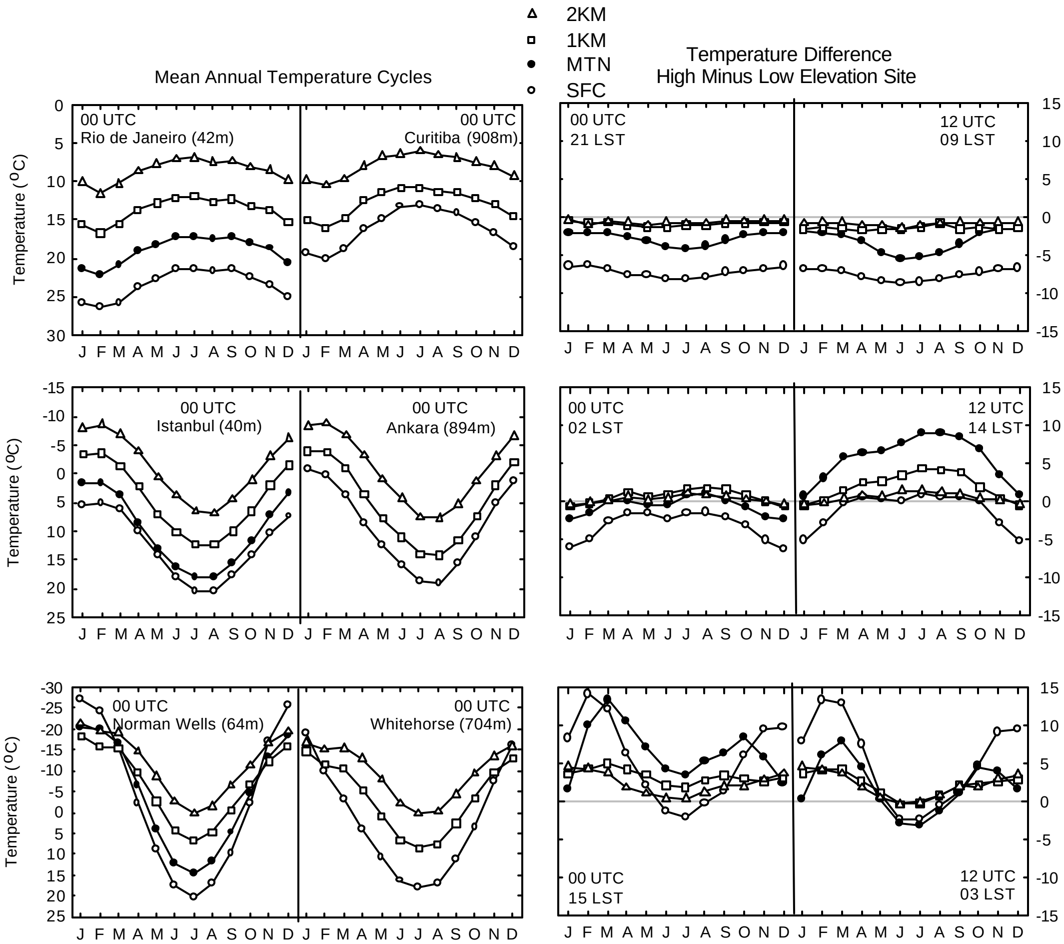
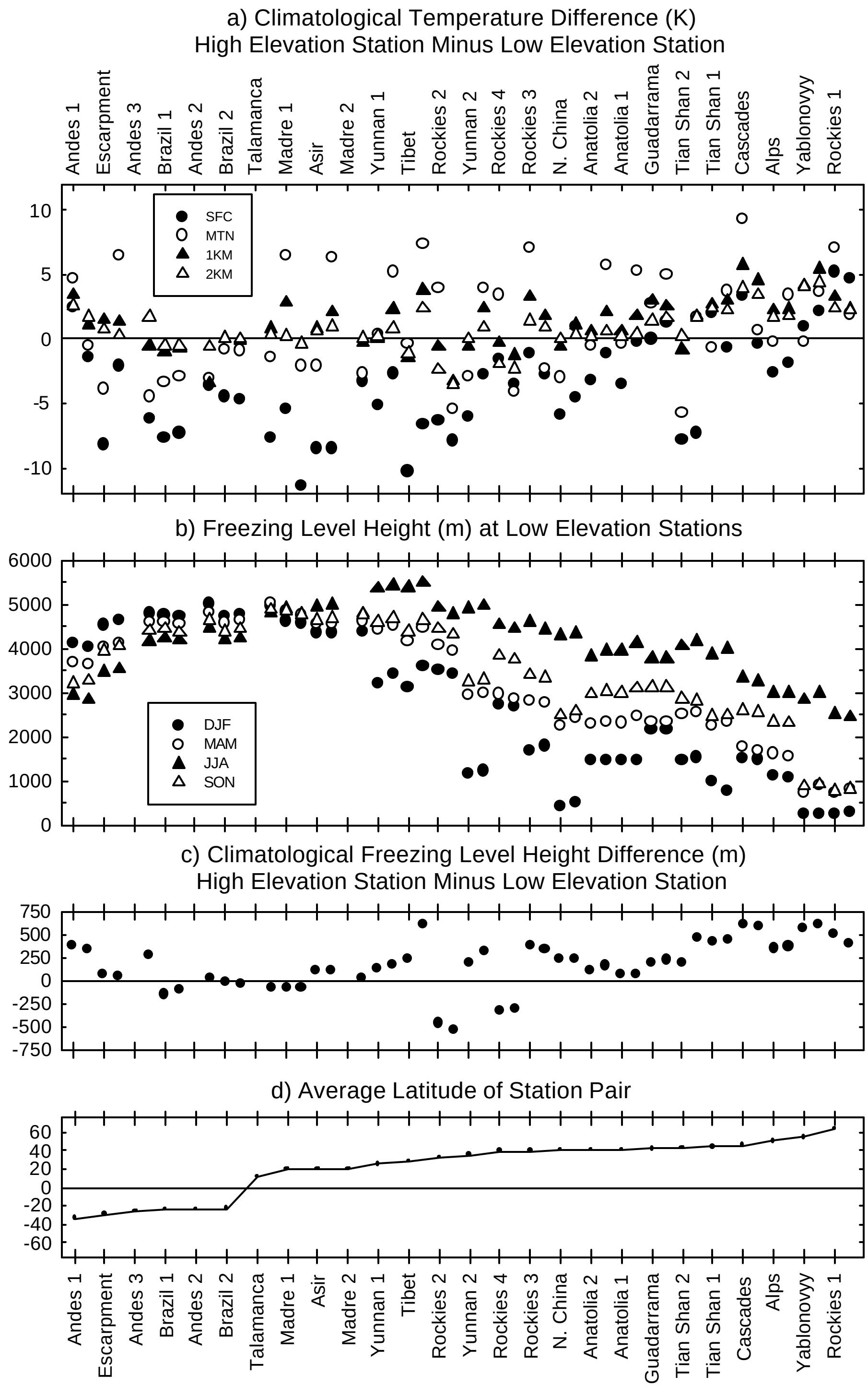

2KM
1KM
MTN
SFC
Mean Annual Temperature Cycles
Temperature Difference
High Minus Low Elevation Site
00 UTC
Rio de Janeiro (42m)
00 UTC
Curitiba (908m)
00 UTC
21 LST
12 UTC
09 LST
00 UTC
Istanbul (40m)
00 UTC
Ankara (894m)
00 UTC
02 LST
12 UTC
14 LST
00 UTC
Norman Wells (64m)
00 UTC
Whitehorse (704m)
00 UTC
15 LST
12 UTC
03 LST
Temperature (^{o}C)
J F M A M J J A S O N D J F M A M J J A S O N D

Figure 3



a) Climatological Temperature Difference (K)
High Elevation Station Minus Low Elevation Station
SFC
MTN
1KM
2KM
b) Freezing Level Height (m) at Low Elevation Stations
DJF
MAM
JJA
SON
c) Climatological Freezing Level Height Difference (m)
High Elevation Station Minus Low Elevation Station
d) Average Latitude of Station Pair
Andes 1
Escarpment
Andes 3
Brazil 1
Andes 2
Brazil 2
Talamanca
Madre 1
Asir
Madre 2
Yunnan 1
Tibet
Rockies 2
Yunnan 2
Rockies 4
Rockies 3
N. China
Anatolia 2
Anatolia 1
Guadarrama
Tian Shan 2
Tian Shan 1
Cascades
Alps
Yablonovyy
Rockies 1

Figure 4

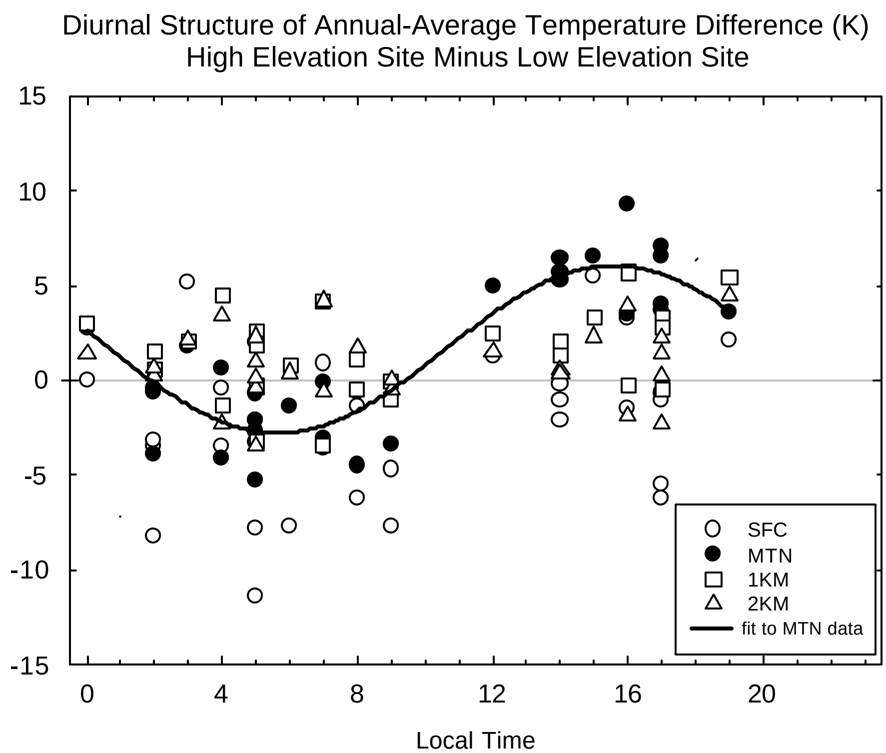

Diurnal Structure of Annual-Average Temperature Difference (K)
High Elevation Site Minus Low Elevation Site
15
10
5
0
-5
-10
-15
0
4
8
12
16
20
Local Time
SFC
MTN
1KM
2KM
fit to MTN data

Figure 5

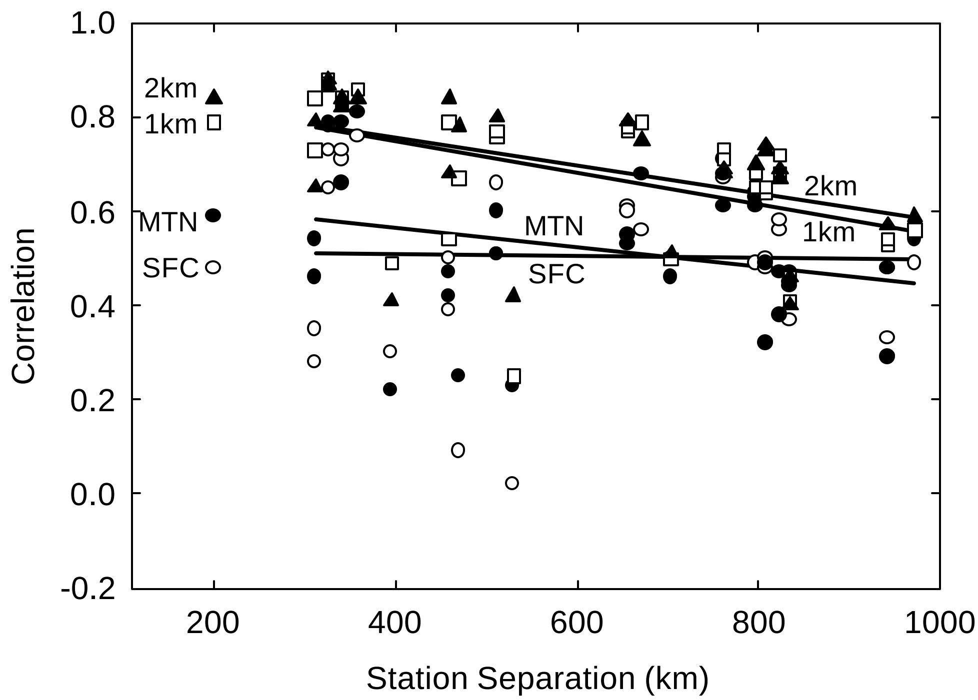

Correlation
Station Separation (km)
1.0
0.8
0.6
0.4
0.2
0.0
-0.2
200
400
600
800
1000
2km
1km
MTN
SFC
2km
1km
MTN
SFC

Figure 6

Temperature Trends (K/decade) vs. Latitude

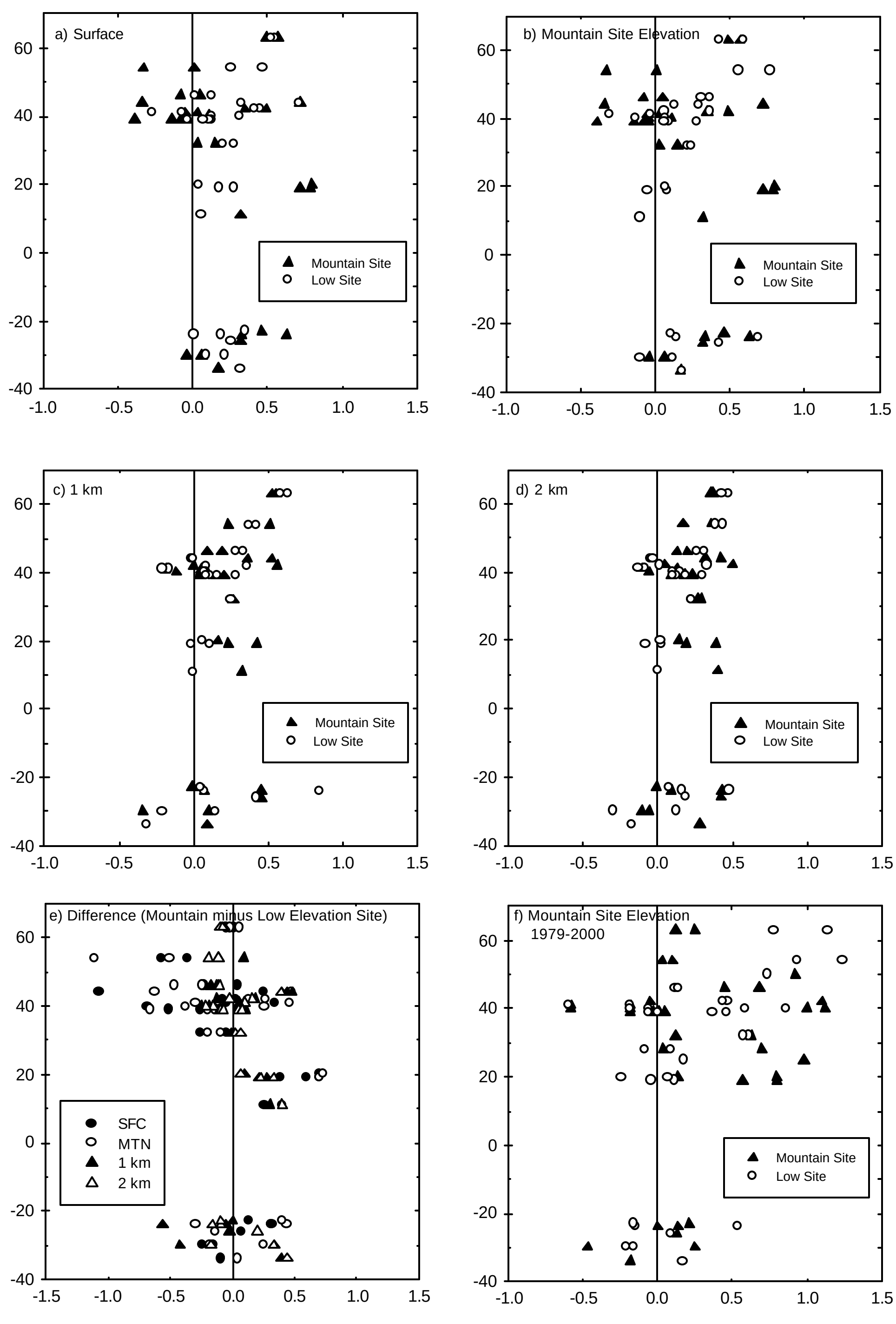


Figure 7

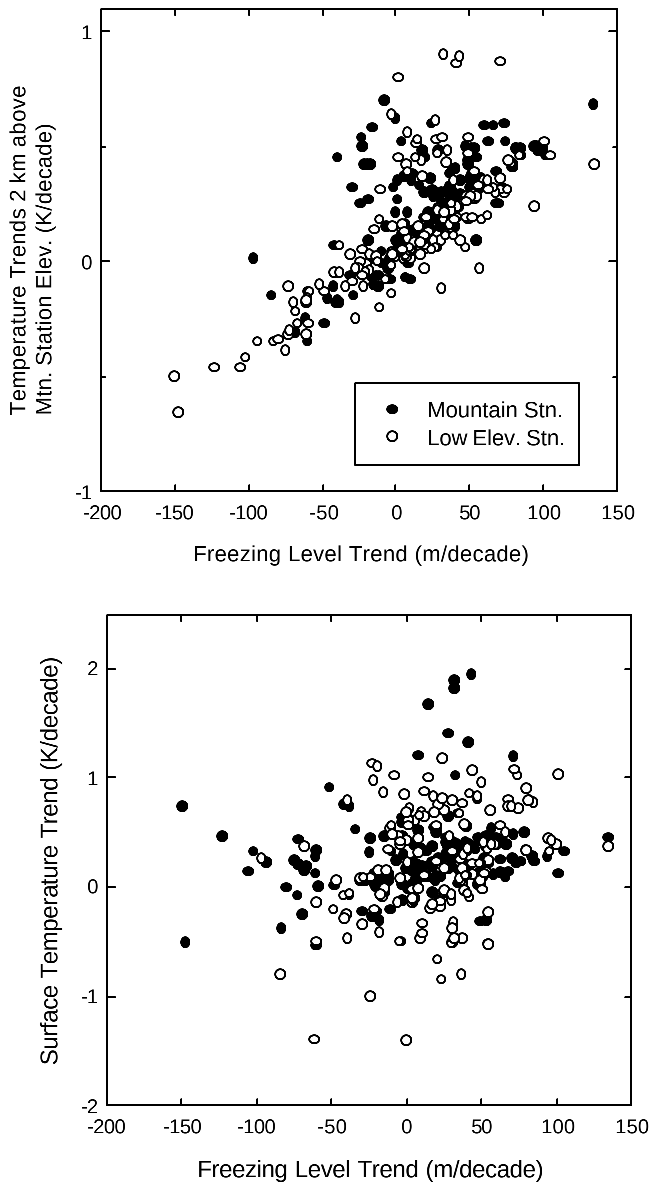

Temperature Trends 2 km above
Mtn. Station Elev. (K/decade)
Mountain Stn.
Low Elev. Stn.
Freezing Level Trend (m/decade)
Surface Temperature Trend (K/decade)
Freezing Level Trend (m/decade)

Figure 8