The Effect of Moisture on Layer Thicknesses Used to Monitor Global Temperatures

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(Manuscript received 31 December 1992, in final form 2 June 1993)

ABSTRACT

Mean layer virtual temperature estimates, based on geopotential height measurements, form the basis for one approach being used to monitor changes in upper-air temperature. However, virtual temperature is a function of atmospheric moisture content as well as temperature. This paper investigates the impact of real or apparent changes in atmospheric moisture on changes in mean layer virtual temperature. Real changes in mean layer specific humidity of up to 50% would cause changes in mean layer virtual temperature of less than 1°C, except in the tropical boundary layer, where the high moisture content would lead to larger virtual temperature changes. The effect of humidity changes is negligible in polar regions and most pronounced in the tropics, which could influence the interpretation of the latitudinal gradient of virtual temperature trend estimates. Improvements in radiosonde humidity sensors since 1958 have led to an apparent decrease in atmospheric humidity. On global average, for the 850–300-mb layer, such changes are estimated to contribute to an apparent cooling of between 0.05° and 0.1°C, or about 10% to 20% of the observed warming trend since 1958.

1. Introduction

Interest in global climate change has led to monitoring of global temperatures, both at the surface and in the upper air. For the latter, the main source of data is the global radiosonde network, from which data adequate for the task are available since about 1958. A convenient measure of the mean temperature in a layer bounded by two predetermined pressures is the geopotential height difference between these two pressure surfaces (the thickness of the layer), which depends on the density of the air in the layer. The density is largely a function of the mean temperature in the layer but also depends on the moisture content. This dependence on both temperature and moisture is expressed by the virtual temperature, T_v, which is the temperature that dry air would have if it had the same density as moist air at the same pressure.

Computed heights of the pressure surfaces depend on the vertical density distribution between the pressures, not just the density at particular pressure levels, and so the thickness of the layer reflects conditions in the entire layer. Angell (1988) and others (Dronia 1974; Hense et al. 1988) have converted thicknesses to mean layer temperatures to monitor temperatures in the troposphere and lower stratosphere.

The temperature obtained from the thickness is the mean layer virtual temperature, $\langle T_v \rangle$, rather than the mean layer temperature, $\langle T \rangle$, which would be measured by a thermometer. Because $\langle T_v \rangle$ is also a function of the moisture distribution in a layer, any change in moisture could be interpreted as a change in mean temperature of the layer. Because models that show substantial warming of the troposphere with increases in greenhouse gases also show substantial increases in water vapor in the troposphere, it is worthwhile to see how much moisture affects apparent mean layer temperatures calculated from thicknesses. We also examine the effect of improved humidity instruments on $\langle T_v \rangle$ estimates.

2. Virtual temperature

In practice, one obtains $\langle T_v \rangle$ by inverting the hypsometric equation for the differences in height between two pressures:

$$\langle T_v \rangle = (Z_2 - Z_1)/[(R_d/g_0)\ln(p_1/p_2)], \quad (1)$$

where $\langle T_v \rangle$ is the mean virtual temperature between pressures p_2 and p_1 at geopotential heights Z_2 and Z_1, respectively; R_d is the gas constant for dry air; and g_0 is the acceleration of gravity at sea level. The equation for T_v at a point is

$$T_v = (1 + 0.608q)T, \quad (2)$$

where q is the specific humidity in grams per gram (Iribarne and Godson 1981).

Usually T_v exceeds T by less than 1°C but in tropical regions with high temperatures and moisture content,

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the difference can be as large as 2°C. At low temperatures, the amount of moisture is so small that T_v and T have nearly the same values. For instance, for saturated air at 300 mb and $T = -45°C$ (approximately the 1976 *U.S. Standard Atmosphere* temperature for midlatitude), T_v would be only 0.04°C larger than T. (At 1000 mb, $T = -33°C$ would produce the same difference.) Above about 300 mb, the effect of moisture on calculations of T_v can be neglected. Below this level, however, the possibility exists that secular changes in moisture could be confused with changes in temperature or could obscure a true change.

There is an approximation to T_v that is often made and will be useful in section 4. Because T_v and T in Eq. (2) are in K, and the range of T is relatively small,

$$T_v \approx T + A \times q \tag{3}$$

where, if q is now expressed in grams per kilogram, A is between 0.16 K and 0.18 K depending on T; below we use 0.17 K, which is appropriate for $T \approx 273$ K.

3. Precision of thicknesses

Before treating moisture effects explicitly we examine the precision of the $\langle T_v \rangle$ measurements based on radiosonde observations of pressure and geopotential height. Because the thickness of the layers depends on moisture and temperature, one might expect the precision of $\langle T_v \rangle$ to be less than that of $\langle T \rangle$. To investigate this we performed a Monte Carlo analysis, using known statistical properties of radiosonde geopotential height measurements.

Equation (1) can be written

$$\langle T_v \rangle = [(Z_2 + \epsilon_{Z2}) - (Z_1 + \epsilon_{Z1})]/[(R_d/g_0)\ln(p_1/p_2)], \tag{4}$$

where ϵ_{Z2} and ϵ_{Z1} are random measurement errors in geopotential heights corresponding to pressures p_2 and p_1. These errors can be simulated as Gaussian random variables with zero mean and standard deviations equal to the rms errors of radiosonde measurements, which range from about 0.4 gpm at 1000 mb to about 9 gpm at 300 mb (National Weather Service 1990). However, it seems reasonable that some positive correlation exists between ϵ_{Z2} and ϵ_{Z1} in a given sounding (Hooper 1975). In this case, errors can be simulated as above but with the constraint that the two errors always have the same sign. Simulations for both cases—random errors and random errors of the same sign—were carried out for atmospheric layers between 1000 and 300 mb, using *U.S. Standard Atmosphere* values for Z_1 and Z_2. In each case, 1000 simulations were performed and the rms error in $\langle T_v \rangle$ was determined.

The results are displayed in Table 1. The rms errors range from 0.3° to 0.7°C, for the random error case, depending on the relative sizes of the rms geopotential height errors. When height errors are constrained to be of the same sign, the $\langle T_v \rangle$ errors are smaller, generally about 0.25°C for the layers simulated, and show no great dependence on thickness. The difference between the two cases is greatest in shallower layers. These values can be compared to the precisions of radiosonde temperature measurements. The rms errors of about 0.6° to 0.7°C for $\langle T_v \rangle$ of the shallower layers, for the random error case, are about twice the estimated precision of individual temperature measurements [0.3°C, National Weather Service (1990)]. On the other hand, the errors in $\langle T_v \rangle$ for the thicker layers, for example, 850–300 mb, and for the same-sign case for all layers are between 0.2° and 0.3°C, comparable to those of individual temperature measurements. Thus the precision of $\langle T_v \rangle$ is probably no worse than that of $\langle T \rangle$ if one monitors thick-enough layers.

TABLE 1. Precisions of mean layer virtual temperature estimates based on geopotential height data.

	rms error (°C)	
Layer endpoints (mb)	Random	Same sign
1000–850	0.31	0.21
1000–300	0.25	0.24
850–700	0.57	0.27
850–300	0.30	0.25
700–500	0.59	0.21
500–300	0.69	0.26

4. The effect of changes in moisture on changes in thickness

It remains possible that changes in moisture in the troposphere could introduce confusion if interpreted as true temperature changes where none were present or mask such temperature changes if, say, drying accompanied warming.

From Eq. (3), a change in $\langle T_v \rangle$, $\delta\langle T_v \rangle$, can be written

$$\delta\langle T_v \rangle = A\delta\langle q \rangle. \tag{5}$$

To obtain an estimate of $\langle q \rangle$, the mean layer specific humidity, note that the precipitable water, W, between pressure levels p_1 and p_2 is

$$W = (1/g_0)\int q dp = (1/g_0)\langle q \rangle(p_1 - p_2) \tag{6}$$

or

$$\langle q \rangle = (1/g_0)W/(p_1 - p_2). \tag{7}$$

Finally, if we express a change in W as a fraction, r, of W, we can write, using $A = 0.17$ K, $g_0 = 981$ cm s^{-2}, with W in centimeters and p in millibars,

$$\delta\langle T_v \rangle = 167Wr/(p_1 - p_2). \tag{8}$$

TABLE 2. Change in mean layer virtual temperature (°C) for a given fractional change, r, in precipitable water, W.

Layer endpoints (mb)	W (cm)	r 0.10	0.25	0.50
	Global			
1000–850	1.25	0.14	0.35	0.70
1000–300	2.55	0.06	0.15	0.30
850–500	1.15	0.05	0.14	0.27
850–300	1.30	0.04	0.10	0.20
700–500	0.45	0.04	0.10	0.19
	Polar			
1000–850	0.30	0.03	0.08	0.17
1000–300	0.70	0.02	0.04	0.08
850–500	0.35	0.02	0.04	0.08
850–300	0.40	0.01	0.03	0.06
700–500	0.13	0.01	0.03	0.05
	Equatorial			
1000–850	2.40	0.26	0.67	1.33
1000–300	4.85	0.12	0.29	0.58
850–500	2.20	0.10	0.26	0.52
850–300	2.45	0.07	0.19	0.37
700–500	0.95	0.08	0.20	0.40

Table 2 gives values of $\delta\langle T_v \rangle$ that one would expect from a given fractional increase in W in various pressure intervals, calculated from Eq. (8). The W values represent average conditions as estimated in Elliott et al. (1991). The thicker the layer, the less a given percentage change in moisture in that layer affects $\langle T_v \rangle$. Generally the global values are not large, usually <0.1°C for changes in W of 10% or less and even changes of 50% do not produce changes in $\langle T_v \rangle$ approaching 1°C. If such moisture increases result from greenhouse gas increases, the accompanying global average temperature increases would likely be several degrees, significantly greater than the apparent increase due to water vapor.

In the tropics, where moisture is much greater, there can be appreciable effects. Hense et al. (1988) calculate an average trend in W between 700 and 500 mb at four stations in the tropical Pacific equivalent to 0.121 cm decade^{-1} for a 20-year period ending in 1984. A change of this magnitude translates into a $\langle T_v \rangle$ trend of 0.1°C decade^{-1}. [Gaffen et al. (1992) found several tropical stations with post-1972 trends in W from the surface to 500 mb of about 0.3 cm decade^{-1}, which also translates to a temperature trend of 0.1°C decade^{-1} for that layer.] Hense et al. also calculate a trend in $\langle T_v \rangle$ of about 0.46°C decade^{-1} for the 850–200-mb layer. If this value is representative of the shallower 700–500-mb layer, the apparent moisture increase contributed significantly to that trend (but see the next section).

There is a situation where moisture changes could confuse the analysis of temperature change, particularly of the pole-to-tropics temperature gradient. Table 2 gives values typical of regions poleward of 60° and of regions within 10° of the equator. For the same percentage increase in moisture everywhere the equatorial regions would show more "warming" than the polar regions. Most 3D simulations of greenhouse warming show greater warming in the polar regions than the tropics, thus leading to reduced pole-to-equator temperature gradient. On the other hand, the differential moisture effect would tend to increase the pole-to-equator gradient.

5. Effects of instrument changes

There is another source of potential problems, namely, changes in humidity sensors in radiosondes, which have occurred in the past and will likely occur in the future. Improvements in sensors generally lead to faster responses to changing humidity and therefore, since humidity usually decreases with height, to lower values in the mean. Some changes we have found could lead to substantial apparent changes in $\langle T_v \rangle$. Typical radiosonde humidity sensors in the 1960s were lithium chloride strips, specially treated hairs, or goldbeater's skin. To a large extent these have been replaced by carbon-based hygristors and thin-film capacitors, which have much better response times. Figures 1 and 2 show examples of the effects of these changes. The upper parts of the figures show monthly anomalies at 500 mb and the lower parts show mean annual values at three pressure levels. At Jeddah, Saudi Arabia (Fig. 1), the

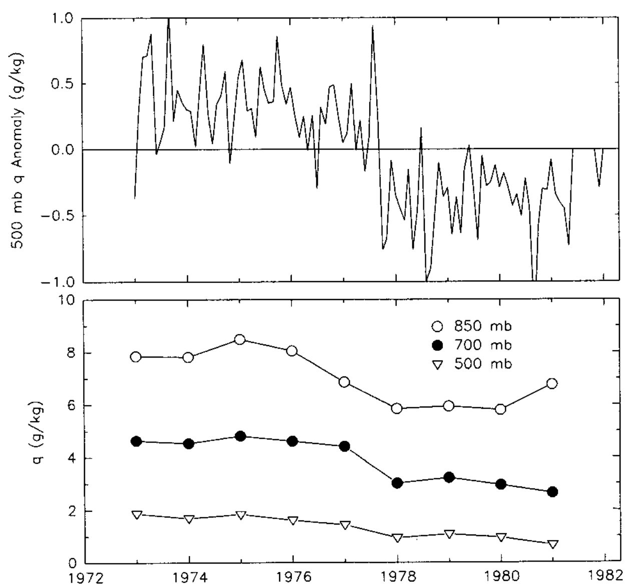

FIG. 1. Monthly specific humidity, q, anomalies at 500 mb (top) and mean annual q at 850, 700, and 500 mb (bottom) at Jeddah, Saudi Arabia (22°N, 39°E), for 1973–81.

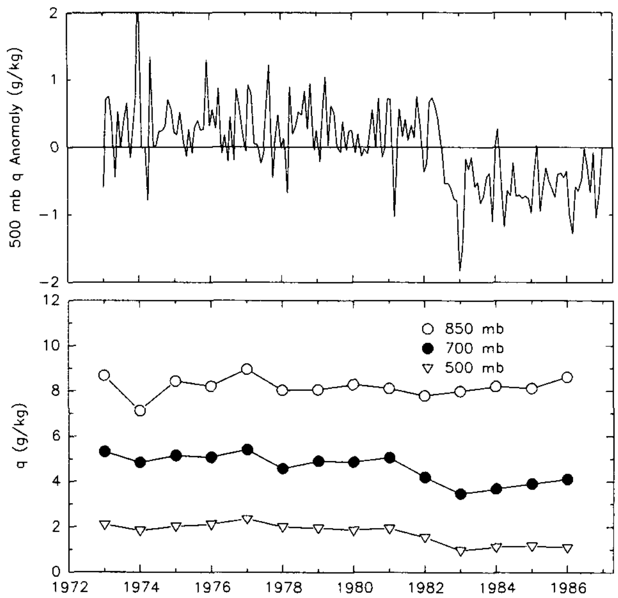

FIG. 2. Monthly specific humidity, q, anomalies at 500 mb (top) and mean annual q at 850, 700, and 500 mb (bottom) for Townsville, Australia (19°S, 147°E), for 1973–86. Note change in vertical scale in top part.

change from hair hygrometer to thin-film capacitive sensor in 1978 accompanied an apparent "drying" of 2.0, 1.5, and 0.8 g kg^{-1}, respectively, at 850, 700, and 500 mb. These translate into apparent decreases in $\langle T_v \rangle$ of about 0.3°C for the 850–700-mb layer and about 0.2°C for the 700–500-mb layer. At Townsville, Australia (Fig. 2), a change from lithium chloride to a carbon-based element in 1982 also shows appreciable apparent drying; at 700 and 500 mb the drop is comparable to that at Jeddah. At higher-latitude stations, with colder temperatures and hence less moisture, the same instrument changes do not show such dramatic moisture discontinuities.

The change at Jeddah is the largest we found in examining some 60 stations (almost all of which are in Angell's network, except for Jeddah) for the period 1973–90. More typical values of $\delta\langle T_v \rangle$ accompanying changes of instruments were 0.1°C or less. Nevertheless, the change at Jeddah may well not be unique during the period since the IGY in 1957. There have been other changes in radiosonde measurements that could affect the calculations of moisture content. The processing of the electrical signals from the sondes has been changed in some cases. Temperature sensors and even the mounting of the sensors can affect the calculation of pressure heights.

An example of an improvement that led to an increase in reported humidity is the effect of changing the housing, discussed by Elliott and Gaffen (1991). They found an apparent rise in daytime relative humidity at Hilo, Hawaii, of about 20% at 850 mb when an improved housing was introduced for U.S. radiosondes. If this same increase occurred at higher levels in the tropics when the new housing was introduced, there would have been an appreciable effect at three of the four stations in the Hense et al. (1988) analysis mentioned above. Such an apparent jump in humidity would lead to an apparent jump in $\langle T_v \rangle$ of close to 0.2°C. Thus a significant amount (over 20%) of the reported average change in virtual temperature (0.92°C) over the 20-year period at the three American stations could have been due solely to improved humidity sensing.

6. Changes in the 850–300-mb layer

Because Angell (1988) has used changes in the thickness between the 850- and 300-mb levels as a measure of changes in tropospheric temperature since 1958, we examined possible effects of instrument changes during this period on this $\langle T_v \rangle$ record, based on a chronology of radiosonde changes from some 50 countries (Gaffen 1993). Instrument changes reported by the weather services often coincided with sharp changes in the humidity records, which were relatively easy to identify. Using this information and examining time series of humidity for most of the stations in the Angell network (since 1973, when our data record begins), we estimated the effects of changes from the slower-responding humidity sensors to more rapidly responding sensors. Averaging q for about a year before and after the change gives a measure of the effect of the change at that station.

Not all countries in the Angell network have reported their history of radiosonde changes, and a number of countries made significant sensor changes between 1958 and 1973 (the United States, for instance). To estimate the effect such changes could have on the record, we assumed that all countries changed from relatively slower- to faster-responding sensors some time during the period. Mean values for the estimated specific humidity decreases from sensor improvements were converted to changes in T_v at 850, 700, and 500 mb. The effect of these assumed changes on the $\langle T_v \rangle$ in the 850–300-mb layer was calculated by mass weighting the changes at those levels and assuming no change in T_v at 300 mb. The latter is justified because, while sensor changes can bring about substantial changes in relative humidity and dewpoint, they do not produce significant specific humidity changes at very low temperatures because the amount of moisture is so small.

Taking representative values of changes in $\langle T_v \rangle$, we found virtually no effect in the polar zone (latitude > 60°), again because of the low temperature. Between 60° and 30°, in both hemispheres, the effect on Angell's temperatures would be a lowering of about 0.05°C between 1958 and the present, and in the tropics the effect would be about 0.09°C. If we take the maximum values for abrupt changes in our record (Jeddah, for the tropics, although it is not part of Angell's network) we estimate 0.01°C in the polar zone, 0.08°C in temperate latitudes, and 0.19°C in the tropics. These values would give an apparent decrease in mean global temperature of 0.06°C for the representative case and 0.13°C for the extreme case. Probably the best way to summarize these calculations is to say that improvements in humidity sensors would have suggested a cooling of the global troposphere somewhere between 0.05° and 0.1°C between 1958 and the present, were there no other changes in the atmosphere or its measurement. Thus the increase of about 0.45°C in the troposphere since 1958 shown in Angell's record could have been between 0.50° and 0.55°C.

7. Conclusions

In monitoring upper-air temperatures from the thicknesses of pressure layers, the effect of moisture on the mean layer temperatures has generally been ignored. The convenience of the pressure height data as a measure of the mean temperature has outweighed the slight error introduced by the moisture. This is particularly true when deviations from a mean have been used, because the mean thickness of a layer incorporates the mean moisture conditions there. Nevertheless, it seemed advisable to quantify the effect of possible changes in moisture on the record since substantial increases in tropospheric water vapor are believed to be a concomitance to global warming. Furthermore, there have been substantial changes in humidity instruments, which could have produced apparent changes in temperatures calculated from thickness data.

Potential increases in water vapor over the next century from increases in greenhouse gases will not distort the global average mean layer temperatures calculated from thicknesses so as to mislead to any great extent. The effect is not negligible, however, especially in the tropics. There is the possibility that the apparent latitudinal gradient of temperature could be affected because moisture increases near the equator would be largest, whereas temperature increases are expected to be smallest there.

Finally, since the improvements in radiosonde humidity devices have generally led to an apparent drying of the troposphere, there would be an apparent cooling. We estimate this effect to be between 0.05° and 0.1°C in the 850–300-mb layer. Angell's record for this layer has an increase of about 0.45°C over 34 years so the actual increase since 1958 could have been between 0.50 and 0.55°C.

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