NOTES AND CORRESPONDENCE

Deriving Significant-Level Geopotentials from Radiosonde Reports

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ABSTRACT

An algorithm is developed to derive hydrostatically balanced geopotentials at significant levels from radiosonde reports of significant-level temperatures and mandatory-level geopotentials and temperatures. It minimizes the square of the nonhydrostatic differences in a layer where at least one significant-level datum is reported and can be viewed as being a 1D analysis step that returns an estimate of the departures from hydrostatic balance within the layer. The piecewise-polynomial interpolation of the minimization procedure is used to produce an expanded geopotential profile in any layer where significant-level data are reported, and the integrated minimization error can be used as a quality-control measure. The algorithm's performance has been evaluated using the global radiosonde dataset for a given synoptic time, and it is found that it produces equivalent layer-mean temperature errors that are generally smaller than radiosonde observational errors.

1. Introduction

During the development of the Canadian regional data assimilation system (Chouinard et al. 1994) it was decided that the system should use significant-level radiosonde data as well as the mandatory-level data used in the previous hemispheric and global systems. The salient features of the processing of radiosonde temperature and, where necessary, moisture data are summarized below and serve to motivate this study.

The onboard instruments measure true temperature and relative humidity during the ascent of a radiosonde, and the measurements are transmitted as a function of decreasing pressure. Detailed profiles of these two quantities are typically available at the station with a vertical resolution of the order of 10 hPa, leading to a computed profile of virtual temperature at the same resolution. The geopotentials at the mandatory levels (1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, and 10 hPa) are computed next. To do so, the detailed profile of virtual temperature is integrated upward using the discretized hydrostatic equation, together with the station pressure and its elevation above mean sea level, as a boundary condition. Significant levels are obtained by applying various selection criteria (World Meteorological Organization 1988) to the full-resolution profiles. The temperatures at the mandatory and the significant levels, the geopotentials at the mandatory levels (but not the significant ones), and the station pressure (the station elevation is available from a station dictionary) are then transmitted on the Global Telecommunications System (GTS).

In principle it should be possible to compute the geopotentials at all levels using only the reported mandatory- and significant-level virtual temperatures and the station pressure and its elevation and, in particular, to reproduce the reported mandatory-level geopotentials. In practice this is impossible since these latter geopotentials are necessarily computed at significantly lower resolution than that used to compute the reported mandatory-level geopotentials, and thus discrepancies will result due to a loss of information. A further consequence is that integrating downward from the nearest mandatory level above a significant level to compute the geopotential there results in a different value from that obtained by integrating upward from the nearest mandatory level that is below it. This discrepancy can be quite significant (particularly in the lower troposphere) and is often compounded by errors that frequently occur in data reporting, transmission, and de-
coding (Gandin 1988; Collins and Gandin 1990), of which data transmission is the most important error source. Indeed, Collins and Gandin noted that 7% of all rawinsonde reports received at the U.S. National Meteorological Center contain at least one hydrostatic error.

Since significant-level geopotentials are absent from the radiosonde reports distributed on the GTS, presumably for historical reasons related to problems of communications bandwidth, the algorithm presented in this note was developed to derive them in a self-consistent manner as possible from the significant-level temperatures and the mandatory-level geopotentials and temperatures. It has been in operational use at the Canadian Meteorological Centre since April 1991 and contributes to the extraction of a maximum of information from this expensive source of data. It can also serve as a component of a complex quality-control system for rawinsonde data. Such systems have been reviewed in Gandin (1988) and a detailed description of a comprehensive one for rawinsonde height and temperature data at mandatory levels is given in Collins and Gandin (1990). The present algorithm could in principle serve as a component for a generalization of the Collins and Gandin comprehensive hydrostatic quality-control system to include significant-level temperature data.

2. Derivation of the algorithm

a. Problem definition

1) Notation

Let \( \Phi = gZ \) be the geopotential and, for notational brevity, let \( T \) represent virtual temperature rather than actual temperature. Also let \( q = \ln(p) \), where \( p \) is pressure, in order to simplify the notation for the hydrostatic relation that can then be written as

\[
\frac{\partial \Phi}{\partial q} = -RT. \tag{2.1}
\]

2) Given and to be determined

Let us assume (see the schematic of Fig. 1) that \( \Phi \) and \( T \) are given at two mandatory levels \( p = p_0 \) and \( p = p_{N+1} \) and that \( T \) is given at the \( N \) intervening significant levels \( p = p_1, p_2, \ldots, p_N \), where \( p_0 < p_1 < p_2 < \cdots < p_N < p_{N+1} \). We seek \( \Phi \) at these \( N \) significant levels.

3) A difficulty

Consider the following discretized version of the hydrostatic equation (2.1):

\[
\frac{\phi_{k+1} - \phi_k}{h_k} = -R \left( \frac{T_k + T_{k+1}}{2} \right), \tag{2.2}
\]

where \( h_k = q_{k+1} - q_k = \ln(p_{k+1}/p_k) \), and subscript \( k \) denotes evaluation at \( p = p_k \). Applying (2.2) to each layer \([p_k, p_{k+1}]\) for \( k = 0, 1, 2, \ldots, N \) yields \((N + 1)\) equations (i.e., one per layer) for the \( N \) unknowns \( \phi_1, \phi_2, \ldots, \phi_N \). There is thus one equation too many, and in general, the problem is overdetermined with no solution (there will only be a solution if one of the equations is fortuitously a linear combination of the others). The reason for this is that the temperature profile varies only approximately linearly in the variable \( q = \ln(p) \) for each layer \([p_k, p_{k+1}]\). In reality, the station observer has further sublayer temperatures available, each one of which satisfies the discrete hydrostatic equation in the sublayer. However, only a subset of the observed temperatures (i.e., those at the mandatory and significant levels) is reported together with computed geopotentials at the mandatory levels. Note that because these geopotentials are computed using (2.2) and \( all \) of the observed temperatures (including many that are not reported), they are generally very accurate.

b. Solution strategy

Since the above system is overconstrained, it motivates relaxing one of the constraints, and the constraint
that $T$ behaves linearly in $q$ in the interval $[p_k, p_{k+1}]$ is the one chosen. We consider and compare the following two ways of doing this (see the schematic of Fig. 2):

(i) Virtual temperature $T$ behaves quadratically in $q$ (i.e., the order of the polynomial is raised but the number of layers remains unchanged).

(ii) The layer is split into two equal sublayers and $T$ varies linearly in $q$ in each sublayer (i.e., the order of the polynomial remains unchanged but the number of layers is doubled).

Note that if the reported significant-level and mandatory-level data are 100% hydrostatically consistent, then both methods reduce to a single straight-line representation of $T$ within the layer $[p_k, p_{k+1}]$; for method (i) the parabola collapses to a straight line, and for method (ii) the two straight-line segments become collinear.

Relaxing the constraint on the variation of $T$ within layers by either of the two methods above results in the system of equations being changed from an overdetermined one to an underdetermined one, and it remains to define precisely how to determine a unique solution. Let $\hat{T}(q)$ be the piecewise-linear [in the variable $q = \ln(p)$] representation between the points $p_0, p_1, p_2, \ldots, p_N, p_{N+1}$ (i.e., the solid lines of Fig. 1), and let $T(q)$ be the profile that varies between the same points in either of the ways (i) and (ii) defined above. Further, let

$$E(q) = T(q) - \hat{T}(q).$$  \hfill (2.3)

To make the solution unique we simply minimize

$$I(\phi_1, \phi_2, \ldots, \phi_N) = \int_{q_0}^{q_{N+1}} [E(q)]^2 dq$$  \hfill (2.4)

with respect to the $N$ parameters $\phi_1, \phi_2, \ldots, \phi_N$ by setting

$$\frac{\partial I}{\partial \phi} = 0$$  \hfill (2.5)

for $v = \phi_1, \phi_2, \ldots, \phi_N$. Note that among a family of temperature profiles this yields the one that is closest to $\hat{T}(q)$ in the least-squares sense. Note also that

$$\bar{E} = \frac{1}{2} \int_{q_0}^{q_{N+1}} [E(q)]^2 dq$$  \hfill (2.6)

is the rms difference between $T(q)$ and $\hat{T}(q)$ in the layer $[p_k, p_{k+1}]$ [it will be identically zero if $T(q)$ and $\hat{T}(q)$ coalesce in each and every interval, see Fig. 2] and is a measure of how well the hydrostatic condition is satisfied by the data. Consequently, it can be used to control the quality of the significant-level temperatures.

c. Representation of $\hat{T}(q)$

Since $\hat{T}(q)$ is assumed piecewise linear [in $q = \ln(p)$] in each layer $[p_k, p_{k+1}]$, it is explicitly written as

$$\hat{T}(q) = T_k \left( \frac{q_{k+1} - q}{h_k} \right) + T_{k+1} \left( \frac{q - q_k}{h_k} \right)$$  \hfill (2.7)

for $k = 0, 1, 2, \ldots, N$.

d. Representation of $T(q)$ for method (i)

To obtain the explicit representation of $T(q)$ using method (i) of section 2b, let $\phi(q)$ vary cubically in $q$ in the layer $[p_k, p_{k+1}]$ such that

$$\phi_k = \phi(q_k), \quad \phi_{k+1} = \phi(q_{k+1}), \quad \frac{\partial \phi}{\partial q}\bigg|_{q_k} = -RT_k,$$

$$\frac{\partial \phi}{\partial q}\bigg|_{q_{k+1}} = -RT_{k+1}. \hfill (2.8)$$
Note that the hydrostatic condition has been introduced here and applied pointwise at levels $p_1, p_2, \ldots, p_N$. Thus,

$$
\phi(q) = \phi_k \left( \frac{q_{k+1} - q}{h_k} \right)^2 \left[ 2(q - q_k) + h_k \right] 
+ \phi_{k+1} \left( \frac{q - q_k}{h_k} \right)^2 \left[ 2(q_{k+1} - q) + h_k \right] 
- RT_k \left( \frac{q_{k+1} - q}{h_k} \right)^2 \left( q - q_k \right) 
+ RT_{k+1} \left( \frac{q - q_k}{h_k} \right)^2 \left( q_{k+1} - q \right),
$$

(2.9)

so

$$
T(q) = -\frac{1}{R} \frac{\partial \phi}{\partial q} 
= -T_k \left( \frac{q_{k+1} - q}{h_k} \right) \left[ 2 \left( \frac{q - q_k}{h_k} \right) - \left( \frac{q_{k+1} - q}{h_k} \right) \right] 
- T_{k+1} \left( \frac{q - q_k}{h_k} \right) \left[ 2 \left( \frac{q_{k+1} - q}{h_k} \right) - \left( \frac{q - q_k}{h_k} \right) \right] 
- \frac{6}{R} \left( \phi_{k+1} - \phi_k \right) \left( \frac{q_{k+1} - q}{h_k} \right) \left( q - q_k \right) 
- \frac{6}{R} \left( \phi_k - \phi_{k+1} \right) \left( \frac{q_{k+1} - q}{h_k} \right) \left( q - q_k \right),
$$

(2.10)

Note also that by construction, $T(q)$ behaves quadratically and passes through $T_k$ and $T_{k+1}$ as formulated above.

e. Representation of $T(q)$ for method (ii)

To obtain the explicit representation of $T(q)$ using method (ii) of section 2b, let

$$
T(q) = 2T_k \left( \frac{q_{k+1/2} - q}{h_k} \right) + 2T_{k+1/2} \left( \frac{q - q_k}{h_k} \right),
$$

for $q \in [q_k, q_{k+1/2}]$,

$$
= 2T_{k+1/2} \left( \frac{q_{k+1} - q}{h_k} \right) + 2T_{k+1} \left( \frac{q - q_{k+1/2}}{h_k} \right),
$$

for $q \in [q_{k+1/2}, q_{k+1}]$, (2.11)

where $q_{k+1/2} = (q_k + q_{k+1})/2$. Integrating this expression for $T(q)$ from $[q_k, q_{k+1/2}]$ and $[q_{k+1/2}, q_{k+1}]$, respectively, gives

$$
\phi_{k+1/2} - \phi_k = -\frac{Rh_k}{4} \left( T_k + T_{k+1/2} \right),$$

$$
\phi_k - \phi_{k+1} = -\frac{Rh_k}{4} \left( T_{k+1/2} + T_{k+1} \right),
$$

(2.12)

and adding these yields

$$
T_{k+1/2} = \left( \frac{T_k + T_{k+1}}{2} \right) - \frac{R}{2} \left( \frac{\phi_{k+1} - \phi_k}{h_k} \right),
$$

(2.13)

Substitution of this expression for $T_{k+1/2}$ into (2.11) gives the explicit representation of $T(q)$ for method (ii). Note that the hydrostatic relation has also been introduced here, but this time in a layerwise manner over half layers [cf. (2.12)] instead of the pointwise way of method (i).

f. Minimization procedure

Using (2.3), (2.7), and (2.10) – (2.11), the lack of hydrostaticity $E_k(q)$ in the layer $[p_k, p_{k+1}]$ may be written as

$$
E_k(q) = f_k(q) \epsilon_k,
$$

(2.14)

where

$$
\epsilon_k = \frac{\phi_{k+1} - \phi_k}{h_k} + R \left( \frac{T_k + T_{k+1}}{2} \right)
$$

(2.15)

is a residual quantity for the layer $[p_k, p_{k+1}]$. For method (i),

$$
f_k(q) = -\frac{6}{R} \left( \frac{q_{k+1} - q}{h_k} \right) \left( q - q_k \right),
$$

(2.16)

whereas for method (ii),

$$
f_k(q) = -\frac{4}{R} \left( \frac{q - q_k}{h_k} \right), \quad \text{for } q \in [q_k, q_{k+1/2}],
$$

$$
= -\frac{4}{R} \left( \frac{q_{k+1} - q}{h_k} \right), \quad \text{for } q \in [q_{k+1/2}, q_{k+1}].
$$

(2.17)

Minimizing $I$ [defined by (2.4)] w.r.t. the parameters $\phi_m$, for $m = 1, 2, \cdots, N$, yields

$$
\left( \frac{\phi_{m+1} - \phi_m}{h_m} \right) - \left( \frac{\phi_m - \phi_{m-1}}{h_{m-1}} \right) = -\frac{R}{2} \left( T_{m+1} - T_{m-1} \right),
$$

for $m = 1, 2, \cdots, N$, (2.18)

where the relation

$$
\int_{q_k}^{q_{k+1}} [f_k(q)]^2 dq = c h_k
$$

(2.19)

has been used, and $c = 6/(5R^2)$ and $4/(3R^2)$, respectively, for methods (i) and (ii). Note that (2.18) holds true irrespective of whether method (i) or (ii) is chosen to represent $T(q)$ within sublayers.

Since $\phi_0$ and $\phi_{N+1}$ are known explicitly (by definition they are the mandatory-level geopotentials), the algorithm reduces to solving the tridiagonal system.
\[
\begin{bmatrix}
-\left(\frac{1}{h_0} + \frac{1}{h_1}\right) & \frac{1}{h_1} \\
\frac{1}{h_1} & -\left(\frac{1}{h_1} + \frac{1}{h_2}\right) & \frac{1}{h_2} \\
\vdots & \ddots & \ddots \\
\frac{1}{h_{N-2}} & -\left(\frac{1}{h_{N-2}} + \frac{1}{h_{N-1}}\right) & \frac{1}{h_{N-1}} \\
\frac{1}{h_{N-1}} & -\left(\frac{1}{h_{N-1}} + \frac{1}{h_N}\right) & \frac{1}{h_N}
\end{bmatrix}
\begin{bmatrix}
\phi_1 \\
\phi_2 \\
\vdots \\
\phi_{N-1} \\
\phi_N
\end{bmatrix}
= \begin{bmatrix}
-\frac{R}{2} (T_2 - T_0) - \frac{\phi_0}{h_0} \\
-\frac{R}{2} (T_3 - T_1) \\
\vdots \\
-\frac{R}{2} (T_N - T_{N-2}) \\
-\frac{R}{2} (T_{N+1} - T_{N-1}) - \frac{\phi_{N+1}}{h_N}
\end{bmatrix}
\] (2.20)

where \( c = 6/(5R^2) \) and \( 4/(3R^2) \), respectively, for methods (i) and (ii). Thus method (i) gives a profile \( T(q) \) that is closer to a piecewise-linear fit to the data than does method (ii), since \( I \) is then smaller. It is therefore preferable to use method (i) when interpolating between significant levels. However, both choices lead to the same values for \( \phi \) at the significant levels.

3. Some results

The minimization algorithm was developed to produce hydrostatically balanced geopotentials at significant levels where only temperature and moisture, but not geopotential, are available. Geopotential is the mass variable of our assimilation system, and it is only through this variable that the significant-level temperature data can enter the analysis system. Because the significant-level data sometimes form very shallow layers (occasionally less than 5 hPa), it is very important to ensure that inconsistencies between temperatures and geopotentials do not lead to large layer-mean hydrostatic errors.

Geopotentials have historically been calculated at significant levels by integrating the hydrostatic equation (2.2) upward from one mandatory level toward the next using the (virtual) temperatures at the significant and the mandatory levels, and the geopotential at the lower of the two mandatory levels. However, as discussed in
the introduction, in general, this will not reproduce the reported geopotential at the uppermost of these two mandatory levels. Since there is in principle no need to integrate all the way up to this level, the integration historically has been stopped at the uppermost significant level, thereby introducing a lack of hydrostatic balance in the layer between the last significant level and the adjacent (uppermost) mandatory level.

This is illustrated in Table 1 where we present data for the lowest two mandatory layers of a radiosonde report. Defining hydrostatic error to be $\varepsilon/R$ [cf. (2.15)] to give units of temperature, the heights ($\phi/\rho$, column 2) derived hydrostatically as described immediately above, by definition satisfy (2.15). They therefore have no hydrostatic error (column 3), except in the uppermost significant layers (925–966 and 850–877 hPa) between mandatory levels, where the errors are $-25.0^\circ$ and $33.0^\circ$C, respectively. The heights (column 5) derived using the minimization algorithm of section 2 have a reduced level of hydrostatic error of $-13.9^\circ$ and $12.2^\circ$C, respectively (see column 6), which although somewhat improved is still unacceptably large.

The sign of the error being opposite and of approximately equal magnitude in each of two adjacent mandatory layers is often an indication that the intervening mandatory-level (925 hPa) report may be in error. As discussed in Collins and Gandin (1990) this is an error that typically results from an interchange of digits in a

**Table 1.** Data for the lowest two mandatory layers (925–1000 and 850–925 hPa) of a radiosonde report as a function of pressure (column 1). Observed quantities are in bold; all other quantities are computed. Vertically offset quantities [cf. Eq. (2.15)] in columns 3 and 6 refer to layers, all others to levels. Quantities in column 2 computed by vertically integrating Eq. (2.2) from the lower mandatory level of each of the two layers bounded by a mandatory level. Quantities in column 5 computed using the minimization algorithm derived in section 2.

<table>
<thead>
<tr>
<th>Pressure (hPa)</th>
<th>Height from Eq. (2.2) (m)</th>
<th>Hydrostatic error ($\varepsilon/R$) ($^\circ$C)</th>
<th>Observed temperature (K)</th>
<th>Height from algorithm (m)</th>
<th>Hydrostatic error ($\varepsilon/R$) ($^\circ$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>850.0</td>
<td>1566.2</td>
<td>33.0</td>
<td>291.9</td>
<td>1536.0</td>
<td>12.2</td>
</tr>
<tr>
<td>877.0</td>
<td>1298.0</td>
<td>0</td>
<td>294.2</td>
<td>1278.9</td>
<td>12.2</td>
</tr>
<tr>
<td>892.0</td>
<td>1152.3</td>
<td>0</td>
<td>292.7</td>
<td>1139.3</td>
<td>12.2</td>
</tr>
<tr>
<td>925.00</td>
<td>840.0</td>
<td></td>
<td>293.2</td>
<td>840.0</td>
<td></td>
</tr>
<tr>
<td>925.00</td>
<td>808.3</td>
<td>$-25.0$</td>
<td>293.2</td>
<td>840.0</td>
<td>$-13.9$</td>
</tr>
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<td>966.00</td>
<td>432.9</td>
<td>0</td>
<td>295.2</td>
<td>466.9</td>
<td>$-13.9$</td>
</tr>
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<td>988.00</td>
<td>237.7</td>
<td>0</td>
<td>292.8</td>
<td>242.6</td>
<td>$-13.9$</td>
</tr>
<tr>
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<td>133.0</td>
<td></td>
<td>294.6</td>
<td>133.0</td>
<td></td>
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**Table 2.** Same as for Table 1 except that the corrected observation at 925 hPa is italicized.

<table>
<thead>
<tr>
<th>Pressure (hPa)</th>
<th>Height from Eq. (2.2) (m)</th>
<th>Hydrostatic error ($\varepsilon/R$) ($^\circ$C)</th>
<th>Observed temperature (K)</th>
<th>Height from algorithm (m)</th>
<th>Hydrostatic error ($\varepsilon/R$) ($^\circ$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>850.0</td>
<td>1530.2</td>
<td>$-6.3$</td>
<td>291.9</td>
<td>1536.0</td>
<td>$-2.3$</td>
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<td>877.0</td>
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<td>294.2</td>
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<td>892.0</td>
<td>1116.3</td>
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<td>1118.8</td>
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<td>804.0</td>
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<td>1.9</td>
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<td>133.0</td>
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report. To test this possibility, the last two digits before the decimal point of the 925-hPa height were interchanged and the value of the 925-hPa height thereby changed from 840 to 804 m. The results, presented in Table 2, show that the layer-mean algorithm-derived hydrostatic error estimates ($\epsilon/R$) are indeed much reduced (to 1.9º and -2.3º, respectively) in the two sets of significant layers. This is a good example of the sensitivity of the algorithm and how it could be used in the data-processing stream to identify problematic layers where the hydrostatic balance is significantly violated.

In order to get a better appreciation of the performance of the algorithm on a larger sample of significant layers, the ensemble mean and variance of the hydrostatic error $\epsilon/R$ have been computed for the full global radiosonde dataset at 0000 UTC 21 October 1993 and are presented as a function of pressure in Fig. 3. The variance only exceeds 0.5ºC² at the 300- and 250-hPa levels and is generally below 0.4ºC² below 400 hPa. This is significantly less than the observational error.

In our current global data assimilation system (Mitchell et al. 1993) the observational error varies between 0.6º and 1.5ºC² depending on the level. Hydrostatic errors this small are a good indication of just how well the algorithm performs in generating significant-level geopotentials.

In Fig. 4 a histogram of the hydrostatic error $\epsilon/R$ in the 700–850-hPa layer is presented for a total of 350 soundings reporting at least one significant-level datum. Note how small the error is and how few of the significant layers have errors exceeding ±0.5ºC. Layers with excessively large errors (arbitrarily defined to be those for which $\epsilon/R$ exceeds ±2ºC) were discarded in the preparation of the ensemble-mean statistics. The number of such rejections was, respectively, 23 and 17 in the 925–1000- and 850–925-hPa layers and less than 3 in all other layers. When compared to the total number of significant layers indicated in Fig. 3, this is less than 1% in all layers with the exception of the 925–1000- and 850–925-hPa layers where it is 13.5% and 9%, respectively. In principle, suspect layers should be flagged and subjected to a complex quality-control system in order to identify and, where possible, correct the errors.

4. Conclusions

An algorithm has been developed to calculate balanced geopotentials at significant levels from mandatory-level information (virtual temperature and geopotential) and virtual temperature at significant levels. This algorithm minimizes the square of the nonhydrostatic difference in each of the layers where at least one significant-level temperature is reported.

The algorithm’s performance was evaluated using the full radiosonde dataset at 0000 UTC 21 October 1993. Over this ensemble of radiosonde reports (734 soundings) at a main synoptic time of observation, the algorithm produced equivalent layer-mean temperature errors of approximately 0.1ºC and a maximum layer-
mean variance of approximately 0.6°C², which is less than the observational errors of radiosonde reports.

The ability of the algorithm to depict errors in radiosonde reports was demonstrated on a typical coding error of an operationally received radiosonde report. The minimization error, which behaves like an analysis error, was effective in identifying layers where the hydrostatic balance was significantly violated. As such, it could be used in the data-processing stream of a data assimilation system to flag dubious reports and transfer them to a comprehensive quality-control system such as that recommended in Collins and Gandin (1990).

The described algorithm has been used in the operational data-processing stream at the Canadian Meteorological Centre since April 1991. In this context it also interpolates the significant-level geopotential data to a regular vertical grid and it has been used to expand the geopotential data onto a 43-level vertical grid for the preparation of the regional analyses of the Comparison of Mesoscale Prediction and Research Experiments Project (COMPARE) (Chouinard et al. 1994).

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REFERENCES


