Some aspects of mesoscale pressure field analysis

J. R. Garratt, CSIRO Division of Atmospheric Research, Aspendale, Victoria, Australia

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Analysis of mean sea level (or any other reference level) pressure fields in pre-frontal and frontal passage situations during summertime in eastern Australia is particularly susceptible to calculation of column temperature required in the pressure reduction process. Over elevated, variable terrain large errors in the pressure field can result, both for mesoscale and larger-scale analysis, and examples are given to illustrate the problem.

Introduction

We describe here features of mesoscale* pressure field (MPF) analysis in the context of frontal studies, based on observations from Phase 2 of the Cold Fronts Research Programme (Smith et al. 1982). Results are shown to have significant implications for sub-synoptic (regional) and synoptic analysis as performed by the Bureau of Meteorology in Australia.

Two features of MPF analysis that have emerged as relevant include:
(a) the use of time-series observations from each station in the mesonetwork; and
(b) the process of reduction of station pressure to a suitable horizontal plane where absolute pressure is required.

In the first, spatial resolution of meso-β features requires a mesonetwork with a mean station spacing ideally ~ 10 km. Even for features 50 km or greater in horizontal dimension that characterise cold fronts and squall lines (Garratt et al. 1984), a spacing of 25 to 50 km is desirable, which becomes impracticable for an area of, say, 300 km square. However in the mesoscale analysis of a specific system such as a cold front or squall line, which can be assumed in quasi-steady state for ~ 1 h and which travels steadily across the network, enhanced resolution can be obtained by incorporation of time-series observations. The method is based on those described by, for example, Fujita (1963) and Barnes (1973) for meso-β and meso-γ (2 to 20 km) analysis.

The second point above concerns the reduction of pressure when this is measured on elevated, variable terrain. Standard practice in regional and synoptic analysis involves reduction to mean sea level (MSL) and, where high altitude stations in particular are involved, the introduction of artificial techniques (WMO 1954). Within Australia the reduction process utilises mean monthly temperature (MMT) for each station, in place of the unknown 'hypothetical column' temperature. This is based partly on the study of Colquhoun (1965) which, it is worth noting, was confined to stations in the altitude range 240 to 580 m. We show in this paper that for frontal occasions in particular, use of the MMT is inappropriate and can lead to considerable error in the analysed pressure field at the meso, regional or synoptic scale.

Pressure measurements

We are concerned here with the analysis of absolute pressure on a horizontal (reference) plane, both at the mesoscale (Fig. 1) and regional scale. Terrain elevation is variable in both, predominantly in the range 0 to 250 m in the former and in the range 0 to 750 m in the latter case. At each station in the mesonetwork, pressure and air temperature are recorded digitally at predetermined intervals of between 1 and 30 minutes, and pressure is reduced at a later date.

*Mesoscale here refers mainly to meso-β (20 to 200 km) but includes meso-α (200 to 2000 km) according to Orlanski (1975). In this paper mesoscale will be used for meso-β, and sub-synoptic or regional scale for meso-α.

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Fig. 1 Station array for Phase 2 of Cold Fronts Research Programme (Grid A), with contours shown in metres. Grid B is the proposed array for Phase 3.
Pressure reduction

We take the simplest case of two stations A and B at elevations $z_{i}$ (take this to be MSL) and $z_{j}$ respectively, as shown in Fig. 2. Station pressures and temperatures are measured as $p_{i, A}, T_{i, A}$ at A and $p_{j, B}, T_{j, B}$ at B. In practice, for a given array of stations, pressures $p_{1}, p_{2}$ etc. are referred to MSL or to the highest station ($z_{h}$ say) in special circumstances. If $z_{r}$ is the reference height, knowledge of the temperature profile $T(z)$ at A, between $z_{r}$ and $z_{i}$, allows $p_{i, A}$ to be inferred precisely (assume here that $T$ is virtual temperature). In the absence of the temperature profile (the usual situation), a best estimate of the mean column temperature is required in order to minimise errors in the reduced pressure $p_{i, A}$ at $z_{r}$. If $z_{r}$ is used as the reference height a hypothetical air column at B, between $z_{r}$ and $z_{j}$ (or MSL), is introduced, but errors in the reduced pressure $p_{j, B}$ at $z_{j}$ due to uncertainties in the mean temperature of this column should be comparable with those in $p_{i, A}$.

If we take $z_{r}$ as the reference level, station A pressure at $z_{j}$ ($p_{j}$ where we have removed the superscript A) is given by

$$
\ln \left( \frac{p_{j}}{p_{r}} \right) = -\beta \frac{z_{r}}{T} dz / T \quad \ldots (1)
$$

with $\beta = g / R$, where we use the hydrostatic equation

$$
\frac{\partial p}{\partial z} = -\rho g \quad \ldots (2)
$$

and the gas equation for moist air,

$$
p = \rho RT. \quad \ldots (3)
$$

In the above $g$ is acceleration due to gravity (9.81 m s^{-2}), $R$ the gas constant for dry air (287 J kg^{-1} K^{-1}), $\rho$ the air density and $T = T_{d} (1 + 0.61 m)$, $T_{d}$ being dry-bulb temperature and m the mixing ratio. In practice $m$ may be replaced by q, the specific humidity, to a high degree of approximation. From Eqn 1 we can define a mean column temperature $\tilde{T}$, so that

$$
\ln p_{j} = \ln p_{r} - \frac{\beta (z_{j} - z_{r})}{T} \quad \ldots (4)
$$

where,

$$
\tilde{T}^{-1} = \frac{1}{\Delta z} \int_{z_{r}}^{z_{j}} dz / T
$$

with $\Delta z = z_{j} - z_{r}$. If $T(z) = T_{s} - \alpha z$, $\alpha$ being a lapse rate and $T_{s}$ the screen temperature, then

$$
\tilde{T} \equiv T_{s} - \frac{\alpha \Delta z}{2} \quad \ldots (5)
$$

to a high degree of approximation. As is well-known, the reduced pressure $p_{r}$ depends upon observed station pressure $p_{r}$, the height difference between station and reference level $\Delta z$ and mean column temperature $\tilde{T}$. Errors induced in $p_{r}$, due to uncertainties in quantities on the right-hand side of Eqn 4, are given by

$$
\delta p_{2} = \left( \frac{p_{j}}{p_{r}} \right) \delta p_{1} - \frac{\beta p_{j} \delta z}{\tilde{T}} + \frac{\beta p_{j} \Delta z}{\tilde{T}^{2}} \delta \tilde{T}. \quad \ldots (6(a))
$$

If $\Delta z \leq 1000$ m (the case here), $\Delta p \leq 100$ mb and, to within 5 to 10 per cent uncertainty, we can write $p_{1} \equiv p_{2}$. Calling $p = 0.5 \left( p_{1} + p_{r} \right)$ we can rewrite Eqn 6(a) as,

$$
\delta p_{\text{REF}} = \delta p_{\text{STAT}} - \frac{\beta p}{\tilde{T}} \delta z + \frac{\beta p \Delta z}{\tilde{T}^{2}} \delta \tilde{T}. \quad \ldots (6(b))
$$

where we introduce a general reference pressure $p_{\text{REF}}$ (to replace $p_{r}$) and station pressure $p_{\text{STAT}}$ (to replace $p_{r}$). Equation 6(b) should be equally valid for an elevated plane or MSL.

Fig. 2 Schematic cross-section of elevated, sloping terrain showing for simplicity two stations at A and B.

**Errors in observed pressure**

In the Phase 2 experiment all eleven field barometers were calibrated against two portable aneroid barometers, themselves extensively calibrated with reference to a sub-standard Bureau of Meteorology mercury-in-glass wall barometer. Least-squares fits to MSL pressure fields during non-frontal conditions, and inter comparisons using mean-monthly pressures, revealed a bias of $-0.45$ mb of unknown origin for the station 3 (Mt Gambier) barometer. Otherwise observing errors of less than 0.2 mb were indicated.

**Errors in altitude estimates**

We take $p = 1000$ mb and $\tilde{T} = 288$ K, whence from Eqn 6(b),

$$
\delta p = - \left[ \frac{\beta p}{\tilde{T}} \right] \delta z = -0.12 \delta z \text{ mb}.
$$

Values of $z$ were based on local 'benchmarks' giving $\delta z = \pm 1$ m, whence $\delta p \equiv \pm 0.12$ mb.

**Errors in column temperature**

From Eqn 6(b) we have

$$
\delta p = \left[ \frac{\beta p \Delta z}{\tilde{T}^{2}} \right] \delta T \equiv 0.0004 \Delta z \delta T \text{ mb},
$$

illustrating the well-known result that as column depth $\Delta z$ increases, uncertainties in $\tilde{T}$ become in-
creasingly important. In our analysis we consider two values of Δz corresponding to station elevations of z_s = 250 and 750 m (these are representative of near-maximum elevations in the mesonetwork and Bureau's regional network), relevant to reduction to MSL. In addition we pause to estimate errors in p due to neglect of water vapour, viz. interpreting $\bar{T}$ as dry bulb, not virtual temperature. Now

$$\delta \bar{T} \equiv (0.16T_p)\delta q = 0.9K \text{ for } \delta q = 5 \times 10^{-3}.$$  

If a climatological value of q is assumed, then errors in q of $5 \times 10^{-3}$ imply errors in p of $\pm 0.1$ and $0.3 \text{ mb}$ for $z_s = 250$ and 750 m respectively. This is by no means the major source of uncertainty in $\bar{T}$, which rather arises from the lack of knowledge of $T(z)$ (in the real column if $z_s$ is used as reference). At best, 'screen' temperatures ($T_s$) are available with each pressure measurement which can be used in Eqn 5 to estimate $\bar{T}$. Alternatively current Bureau practice can be adopted and the MMT used for $T_s$, but in frontal situations, when the actual $\bar{T}$ may exceed the MMT by in excess of 20 K, this will obviously lead to erroneous pressures.

To take this further we consider three likely prefrontal situations, as follows:
(a) daytime clear skies;
(b) cloudy (day and night); and
(c) night-time clear skies,
and evaluate model profiles of $T$ or $\theta (\theta = T + 0.0098z)$, where $\theta$ is strictly the potential temperature only when $p = 1000 \text{ mb}$ at $z = 0$ for any one station. From these, mean column temperatures have been deduced in the form $\bar{T} - T_q$ and are referred to as model or actual values. Details are given in Appendix 1 and results shown in Fig. 3.

We assess two methods of pressure reduction: $\bar{T}$ is estimated from the MMT; and $\bar{T}$ is estimated from observed screen temperature. In the latter we assume initially a lapse rate $\alpha$ of 0.0065 K m\(^{-1}\) based on the ICAN standard atmosphere, though $\alpha$ can be allowed to vary according to (a) to (c), such that the mean difference between model and estimated (Eqn 5) values of $\bar{T}$ is approximately zero. This requires:
(a) $\alpha = 0.0161 \text{ K m}^{-1}$;
(b) $\alpha = 0.0098 \text{ K m}^{-1}$; and
(c) $\alpha = -0.0023 \text{ K m}^{-1}$.

Results of these values of $T_s - \bar{T}$ are shown in Table 1. The difference between estimated and boundary-layer model column temperatures $\delta \bar{T} = \bar{T}_{est} - \bar{T}_{model}$ and $\delta p$, the induced pressure error, are shown in Table 2. As expected the use of MMT ($\bar{T} \approx 15$ to $20^\circ\text{C}$) when pre-frontal temperatures may reach 35 to $40^\circ\text{C}$ overestimates the MSL pressure by several mb. In contrast use of screen temperatures (as originally used by the Bureau of Meteorology though without the introduction of assumed lapse rates) limits biases to within $\pm 0.3 \text{ mb}$ for the mesonetwork (noting that these are maximum values for $z_s = 250 \text{ m}$) and $\pm 0.5$ to 1 mb for the regional network.

### Table 1. Model (Appendix 1) and estimated (fixed or variable lapse rate $\alpha$) values of $T_s - \bar{T}$. L is the Monin–Obukhov stability length (in metres).

<table>
<thead>
<tr>
<th>$\alpha$ (K m(^{-1}))</th>
<th>$T_s - \bar{T}$ (K)</th>
<th>L (m)</th>
<th>Model $T_s - \bar{T}$ (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$z = 250 \text{ m}$</td>
<td>$z = 750 \text{ m}$</td>
<td>$z = 250 \text{ m}$</td>
</tr>
<tr>
<td>0.0065</td>
<td>0.8</td>
<td>2.4</td>
<td>-12.5</td>
</tr>
<tr>
<td>0.0161</td>
<td>2.0</td>
<td>6.0</td>
<td>-25</td>
</tr>
<tr>
<td>0.0098</td>
<td>1.2</td>
<td>3.7</td>
<td>-50</td>
</tr>
<tr>
<td>-0.0023</td>
<td>-0.3</td>
<td>-0.9</td>
<td>46</td>
</tr>
</tbody>
</table>
Table 2. Systematic error $\delta p$ (mb) induced in a summertime, pre-frontal situation when actual column temperature differs from estimated temperature (using screen temperature and assumed lapse rate $\alpha$ or MMT); here $\delta \bar{T}$ (in K) = estimated minus model column mean temperature. L is the Monin-Obukhov stability length representing daytime and night-time clear sky conditions, and general cloudy conditions ($L = \infty$).

'Fixed $\alpha$' is based on Eqn 5: $\alpha = 0.0065$;

'Variable $\alpha$' uses $= 0.0161$ (L negative); $\alpha = 0.0098$ (L = $\infty$) and $\alpha = -0.0023$ (L positive).

<table>
<thead>
<tr>
<th>L (m)</th>
<th>Screen with 'fixed $\alpha$'</th>
<th>Screen with 'var $\alpha$'</th>
<th>MMT</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\delta \bar{T}$</td>
<td>$\delta p$</td>
<td>$\delta \bar{T}$</td>
</tr>
<tr>
<td>z = 250 m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-12.5</td>
<td>3.4</td>
<td>-0.3</td>
<td>2.2</td>
</tr>
<tr>
<td>-25</td>
<td>2.4</td>
<td>-0.2</td>
<td>1.2</td>
</tr>
<tr>
<td>-50</td>
<td>1.9</td>
<td>-0.2</td>
<td>0.7</td>
</tr>
<tr>
<td>$\infty$</td>
<td>0.4</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>46</td>
<td>-2.9</td>
<td>0.3</td>
<td>-1.8</td>
</tr>
<tr>
<td>z = 750 m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-12.5</td>
<td>5.0</td>
<td>-1.5</td>
<td>1.4</td>
</tr>
<tr>
<td>-25</td>
<td>3.7</td>
<td>-1.1</td>
<td>0.2</td>
</tr>
<tr>
<td>-50</td>
<td>3.0</td>
<td>-0.9</td>
<td>-0.6</td>
</tr>
<tr>
<td>$\infty$</td>
<td>1.3</td>
<td>-0.4</td>
<td>0.0</td>
</tr>
<tr>
<td>46</td>
<td>-3.3</td>
<td>1.1</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Our results summarised in Table 1 emphasise the artificiality of the MMT method, which should certainly be avoided in pre-frontal and frontal passage situations. Finally we emphasise that reduction of station pressure to an elevated plane is more physically convincing than reduction to MSL since no fictitious column of air is then involved. In practice reduction to MSL should be acceptable so long as uncertainties in column temperatures are not too great, since the elevated screen temperature at $z_2$ is assumed to equal the air temperature at that level as if no terrain existed. According to the profiles in Fig. 3 this will not be the case except in neutral conditions. Otherwise Table 2 and Appendix 1 indicate that differences (in the form of $\theta_e - \theta_{e,0}$) may be up to 4 K, with an average expected value of about 2 K, night or day. Thus to keep pressure errors from this source to less than 0.5 mb, say, a reduction to MSL would require a terrain elevation $\leq 600$ m.

**Mesoscale pressure field analysis**

The mesonetwork for the Phase 2 experiment contained 11 main stations giving p values each 3 minutes at six of these and each 30 minutes generally, suitable for mesoscale analysis. During a frontal event pressure variations at any one station are considerable (Fig. 4) and the implied spatial fields at various times during the frontal passage are of considerable interest. Within the network, station elevations were mostly below 150 m, with station 11 having $z_2 = 295$ m. Pressures were reduced to both MSL and $z = 295$ m, at 30-minute intervals, using actual screen temperatures and $\alpha = 0.0065$ K m$^{-1}$ to estimate $\bar{T}$ on any one occasion from Eqn 5.
exemplars are given of equivalent fields both at MSL and $z = 295$ m. This shows that most of the dominant features at any one time are common to both, and the rotation of the pressure gradient $\nabla p$ between MSL and $z = 295$ m (as evidenced at A) reflects the strong thermal wind at low levels (veering of the geostrophic wind).

### Application to regional analyses

Figure 7 gives two examples of pre-frontal regional analyses for Victoria, during two Events of Phase 2, both at 0400 GMT (1500 local time). The MMT was used in the calculation of MSL pressures, though screen temperatures (and implicitly column temperatures) were 15 to 20 K greater. This fact is almost certainly related to the exaggerated ridge of high pressure across the Great Dividing Range (station elevations up to 1000 m approximately) and to the high pressure cell in the northeast. We have recalculated MSL pressures using screen temperatures, with $\bar{T}$ given by Eqn 5, giving the modified fields shown in Fig. 8. Further details are given in Appendix 2. The main result is the removal of the high pressure ridge across the high terrain, suggesting that this is an artefact of the analysis method (the MMT is arbitrary). Observed winds ($z \approx 10$ m) are also shown and the general impression given by comparison of cross isobar flow between Figs 7 and 8 is one of greater physical consistency in Fig. 8.

Observed winds are more consistent with the pressure analysis in Fig. 8, supporting the argument that the high pressure ridge along the Great Dividing Range (Fig. 7) is an artefact of the original analysis.

### Summary

Mesoscale pressure-field analysis may be performed using MSL or an elevated plane as reference level, and incorporation of weighted time-series observations should be considered. The pressure-reduction method suggests use of actual column temperatures rather than the use of mean monthly temperature in regional and synoptic analysis. The situation will be particularly serious in the pre-frontal and frontal passage situations during summertime over southeastern Australia.

### Acknowledgments

To Sandy Troup and Harvey Stern for preliminary discussions, and to the latter for provision of Bureau of Meteorology regional data and analyses.

### References

- National Severe Storms Laboratory, 60 pp. (NTIS Comm-73-10781).
Fig. 5 Examples of mesopressure fields and radar-echo distribution at 1000 GMT (a) and 1200 GMT (b) during Event 3 of Phase 2 where symbols H1, L1, etc. refer to mesopressure systems. Pressure is in tenths of mb above 990 mb, and wind direction at several stations is shown, with 'C' indicating calm conditions.


Appendix 1
Model calculations of θ
We represent cloudy and clear sky conditions throughout the diurnal cycle by a range of L, the Monin-Obukhov stability length, and appeal to boundary-layer theory to calculate temperature profiles to be expected over the land. Our calculations are based on the following basic assumptions:
(a) daytime boundary-layer depth \( h = 1000 \) m;
(b) surface layer has depth \( h_s = 0.1 \) h and \( 0.2 \) h for \( L = \infty \) or negative and \( L \) positive respectively; and
(c) surface has roughness length \( z_0 = 0.1 \) m and analogous temperature scale \( z_T = 0.01 \) m.
For a cloudy, overcast sky we take \( L = \infty \), i.e. neutral conditions, so that the temperature profile in the boundary layer (zch) is adiabatic and \( θ = θ_{hs} \) (≡ \( θ_{ad} \) say) for \( 0 < z < h \).
In the daytime situation we assume a well-mixed layer between \( h_s \) and \( h \) and take one value of \( L \) to represent partly cloudy (\( L = -50 \) m) and two values
Fig. 6 Mesopressure fields at 0400 GMT during Events 2 (a) and at 1000 GMT during Event 3 (b), calculated at two levels — MSL and $z = 295$ m.

Fig. 7 Simplified Bureau of Meteorology Victorian regional analyses for 0400 GMT on 23/11/81 (a) and 10/12/81 (b). Pressure at each station is in tenths of mb above 1000 mb, and surface wind direction taken from the original analyses are shown as arrows.
Fig. 8 Re-analysis of the two situations of Fig. 7 using original station pressure, observed screen temperature, and Eqn 5.

\[
\theta = \theta_h \quad \text{at} \quad h < z < h_s
\]

\[
\theta - \theta_o = -\frac{\theta_h}{k} \left[ \ln \left( \frac{z}{z_0} \right) - \Psi(z/L) \right] \quad z_s < z < h_s
\]

... A1

where \( \theta_o \) is true surface temperature, \( \theta_s \) the turbulent temperature scale, \( k \) the von Karman constant \((= 0.40)\) and \( \Psi = 2 \ln (1 + y)/2 \) with \( y = (1 - 16z/L)^{1/4} \) (see Appendix I in Garratt and Francy 1978). The stability length \( L \) is defined as \(-u^2/\theta/k \), and we take \( u^2 \), the surface friction velocity, = 0.52 m s\(^{-1}\) as typical for moderately rough terrain and \( \theta = 288 \) K. Equation A1 is then used to calculate \( \theta (z) \). Specifically taking screen height as 2 m, we have

<table>
<thead>
<tr>
<th>Temperature difference (K)</th>
<th>L(m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>L=25</td>
<td></td>
</tr>
<tr>
<td>( \theta_o - \theta_h )</td>
<td>22.1</td>
</tr>
<tr>
<td>( \theta_o - \theta_s )</td>
<td>18.1</td>
</tr>
<tr>
<td>( \theta_o - \theta (z=250) )</td>
<td>21.2</td>
</tr>
<tr>
<td>( \theta_o - \theta (z=750) )</td>
<td>21.8</td>
</tr>
</tbody>
</table>

For the clear sky, night-time situation we take one value of \( L = 46 \) m (the result in Table 1 is not too sensitive to the \( L \) value) based on \( u^2 = 0.25 \) m s\(^{-1}\) and \( \theta_s = -0.1 \) K (typical of values given in Garratt (1982) for three sites), and implying \( h = 136 \) m (see, for example, Garratt (1982) for formulation). For a stable layer \((z < 27.2 \) m), \( \Psi = -5z/L \) so that we again use Eqn A1 to deduce \( \theta (z) \). Between \( h_s \) and \( h \) we take a linear variation in \( \theta \), and assume radiative cooling between \( h \) and \( z = 1000 \) m comparable with that discussed by, for example, Garratt and Brost (1981). At \( h \) itself, the temperature is deduced from a critical Richardson number criterion, viz.

\[
\text{Ri} = 0.2 = \frac{gh (\theta_h - \theta_s)}{\theta G^2}
\]

where \( G \) is the geostrophic wind with \( G^2 = u^2 C_{G-1} \) and we take \( C_{G-1} = 0.5 \times 10^{-3} \) and \( \theta = 288 \) K. We find \( \theta_h - \theta_s = 5.4 \) K, and in addition the following values.

Appendix 2

In the preparation of the regional fields in Fig. 8 special attention was given to the calculation of MSL pressure at two stations.

Bendigo

The two analyses of Fig. 8, plus one other on a day of southerly winds, showed consistently that the mean pressure for four stations in the vicinity was \( 1 \pm 0.3 \) mb lower than that measured at Bendigo. Consequently Bendigo pressure has been reduced by 1.0 mb on the assumption that this is an instrument bias.

Omeo

Most stations in the regional network within Victoria have elevation \( z_s < 300 \) m; one exception is Omeo, with \( z_s = 679 \) m. Use of Eqn 5 then overestimates the probable \( \theta \), if the latter can be assumed that expected for a daytime, clear sky boundary layer as described in Appendix I. On this assumption the probable \( \theta \) will be approximately 3 K less than that calculated, so that we have increased the Omeo MSL pressure by 0.8 mb.