Dynamics of a Rapid Cloud Band Development over Southeastern Australia

G. A. MILLS
Bureau of Meteorology Research Centre, Melbourne, Australia

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ABSTRACT

On the night of 27–28 August 1985, a cloud band associated with a cold front dramatically broadened over southeastern Australia, leading to unforecast rain and overforecast daytime temperatures. This paper presents a detailed synoptic description of this event, describes the vertical circulations leading to cloud band development, and, by means of back trajectory analysis, determines the origins of air parcels entering the cloud band.

The analyses which were used in this diagnostic study were prepared using an intermittent insertion incremental limited-area data assimilation system. The consistency and coherence of the diagnostic quantities calculated from these analyses demonstrates the value of linking the analyses with a dynamic forecast model, even with the relatively coarse resolution of 250 km used in this study.

It is shown that, following the reorganization of the jet patterns over Western Australia (WA), pressure falls lead to an amplification of the surface pressure trough and the acceleration of both the southerly flow west of the trough and the northerly flow east of the trough. The thermal gradient increased over WA in this period under the influence of the increased convergence and increased deformation associated with these flows, which themselves were strongly influenced by the direct vertical transverse circulation at the jet entrance region. Back trajectory analysis shows that air parcels which formed the developing part of the cloud band had their origins in the boundary layer over central Australia, and moved into the ascending warm conveyor belt as a northeasterly baroclinic flow towards the rising branch of the vertical circulation at the jet entrance and thereafter ascended at a rate of up to 330 hPa day⁻¹.

1. Introduction

Many of the significant public weather forecasting problems in southern Australia are associated with cold front passages. For many years researchers have studied the movement and evolution of cold fronts, with the contributions of Berson (1958), Berson et al. (1957, 1959) and Clarke (1961) providing significant early insights, and the series of observational and modeling studies from the Australian Cold Fronts Research Program (e.g., Wilson and Stern 1985; Garratt et al. 1985; Ryan and Wilson 1985; Reeder 1985; Reeder and Smith 1986, 1987; Garratt 1988) being published more recently. Interestingly, most of the published studies have concentrated on summertime cool changes, because of the social and economic impact of misforecast frontal movement leading to major errors in the timing of alleviation of high temperatures, and the potential for severe fire-weather to be associated with these fronts (for example, see Bureau of Meteorology 1984). Another aspect of frontal weather, which can cause major forecast errors, is the rapid thickening of the frontal cloud band as it moves eastwards, leading to unforecast rain or overforecast prefrontal temperatures. Little appears in the literature documenting cases of frontal cloud band development in Australia, although Downey et al. (1979) do describe one case associated with a cutoff low, and a mechanism for the development of observed prefrontal cloud bands is included in the conceptual model of the summertime cold front proposed by Ryan and Wilson (1985).

On 27 August 1985, a cold front was moving through South Australia (SA, see location map, Fig. 1), and during the afternoon and evening of that day, clouds rapidly developed into a band some 10° longitude wide, and by the morning of 28 August rain was falling in a band from central SA to the western half of Victoria. This study uses a series of 250 km horizontal resolution objective analyses prepared at 6-hour intervals, and diagnostics computed from these analyses, to describe in some detail the synoptic evolution of the frontal system as it moves across Australia. The objective analysis system used is a new incremental data assimilation system which is being developed for operational use by the Australian Bureau of Meteorology, and the linking of analyses at 6-hour intervals with a primitive equations forecast model provides a set of synoptic-scale analyses over the Australian region which have hitherto unavailable temporal and dynamic consistency. The limitations of analysis resolution used,

Corresponding author address: Dr. G. A. Mills, Bureau of Meteorology Research Centre, GPO Box 1289K, Melbourne VIC 3001, Australia.

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however, does mean that in the context of this study, the evolution of the broad baroclinic zone, rather than the detailed structure of the frontal transition zone, is the principal feature to be addressed. This is not inappropriate given that it is the development of a major cloud band which is being studied; selected surface observations will be presented, however, to relate the analyses to observed weather.

The following sections of this paper describe the analysis system and data used in the preparation of the analyses, the synoptic evolution of the frontal system as it moves across the continent, and an attempt is made to identify some of the interactions between the jet streak circulations, frontogenesis and ageostrophic flows prior to the development of the cloud band. Finally, by means of back trajectory analysis, parcel trajectories are categorized and the origins of the airflow associated with the developing cloud band are described.

2. Preparation of the analyses

Real data diagnostic studies of the dynamics of frontal systems and their interaction with their environment have followed two paths. The first approach, followed by, for example, Sanders (1955), Bosart (1970) and Ogura and Portis (1982), uses analyses of various combinations of surface, routine radiosonde, and special network radiosonde data, and then calculates diagnostic quantities from the gridded variables. While reflecting the "real" data and providing excellent diagnostic results in these cases, there can be difficulties in applying this method. For example, the observational data density may well be inadequate to resolve the scales being studied unless, like Ogura and Portis, the front happens to be traversing a special observing network. Even then only a small section of the whole front is sampled at the highest observational resolution since special networks, by their nature, tend to be localized. Interpretation of the results can also be hampered by lack of temporal continuity between analyses due to lack of dynamical constraints and variable observational density, and by the well-known errors involved in computing divergence and vertical velocity from upper-wind soundings. This last-mentioned factor has been the reason for many diagnostic studies using (with considerable success) quasi-geostrophic theory to infer vertical motion.

The second approach (e.g., Anthes et al. 1982; Ross and Orlanski 1982; Orlanski and Ross 1984) uses the well-documented ability of meso-α/β scale NWP models to produce realistic mesoscale detail in their forecast fields and to obtain dynamically consistent time series of gridded data. Most such simulations are initialized from synoptic-scale analyses, with a guess field obtained from a large-scale analysis or forecast. Keyser and Uccellini (1987) have extensively reviewed this approach to diagnostic synoptic meteorology and discussed how it can complement purely observational studies.

In this paper the analyses used may not necessarily overcome all the deficiencies of either approach, but do combine some of the benefits of these two methods. They are prepared by a limited-area incremental data assimilation scheme, which is being developed for operational use in the Australian Bureau of Meteorology and which was operated in test mode using data from 0000 UTC 15 August 1985 to 0000 UTC 31 August 1985. The analysis system consists of two main components. The first is the analysis, or correction part, where a guess field is corrected to fit the observed data, and a forecast component, whereby the updated guess field (which comes from a previous forecast) is then integrated to the next analysis time. In the configuration used in this study, a 6-hour update cycle was used. Thus at each 6-hour analysis time, the analyses reflect the observations, but have a strong dynamic constraint provided by the forecast model.

During the analysis (correction) phase of the data assimilation cycle, deviations of observations from the forecast (interpolated to pressure coordinates) are analyzed on pressure surfaces (in this case at nine levels: 1000, 850, 700, 500, 300, 250, 200, 100 and 50 hPa) on a 250 km grid over the Australian operational limited-area forecast domain (for example, see Fig. 2 of Leslie et al. 1985). Fields are analyzed in the order of mean-sea-level pressure (MSLP), geopotential layer thickness, orthogonal wind components, temperature and dewpoint depression. MSLP and geopotential layer thickness are analyzed using a two-dimensional univariate statistical interpolation (SI) scheme, wind components using a three-dimensional SI scheme, and temperature and dewpoint using a successive correction method (SCM) scheme.
The data used were extracted from the Bureau of Meteorology (B of M) archives of incoming coded observations from the local communications network and the Global Telecommunications System (GTS). Surface data consisted of SYNOPs, ships and buoys. Tropospheric analyses used radiosonde and radar wind observations, Tiros Operational Vertical Sounding (TOVS) data processes by NESDIS in Washington, DC, and transmitted via the GTS at approximately 500 km resolution, cloud-drift winds processed in Japan from the GMS-3 geostationary satellite, and single-level winds from aircraft. As the observational network of ships and drifting buoys is inadequate to resolve the surface pressure pattern over the ocean, operational analyses in the National Meteorological Centre (NMC) of the B of M use bogus observations, termed PAOBS, which are generated by the techniques described by Guymer (1978). The PAOBS for the operational hemispheric analyses are archived, and were retrieved for use in these analyses. Also during this period, one of the orbiting satellites was producing microwave-only soundings, and thus did not produce moisture profiles.

Figure 2 shows typical data distributions for surface and upper air data. Single level data are not shown; cloud-drift wind data from the GMS-3 satellite are concentrated over the Indian and Pacific oceans north of 35°S in the low levels, and are scattered more widely and sparsely in the upper levels. The average number of observations available at each analysis time over the period 0000 UTC 15 August to 0000 UTC 31 August 1985 is shown in Table 1. The tropospheric analysis is primarily dependent on TOVS data over the ocean areas, and due to the sparsity of radiosondes at other than 0000 UTC is also heavily dependent on TOVS and the upper-wind network data over land at other times. TOVS data below 700 hPa are not used over land due to potential bias problems in these statistical retrievals near land surfaces.

The analysis system is designed with the data availability in mind; the guess fields (the 6-hour forecast fields) are adjusted after each prior analysis step to ensure maximum consistency in the analyses. For example, after the MSLP analysis the guess-field geopotential fields are adjusted using a vertical correlation function to distribute a hydrostatically calculated surface geopotential increment. After the thickness analyses, the wind component guess fields are adjusted by a geostrophic increment before the wind component analyses are done, and the temperature guess fields are adjusted hydrostatically to reflect the analysis thickness increments. In the preparation of these analyses temperature fields were computed from the analyzed thickness fields, rather than perform separate SCM analyses, as this was found to produce slightly more accurate 6-hour forecasts.

To make maximum use of the preponderance of mass data over the oceans, a variational merging of geopotential increment and wind component increment analyses is performed, using the method described by Seaman et al. (1977) to derive adjusted mass and wind component increment fields which mutually reflect the other, with the relative weights being dependent on analysis error variances carried through from the SI analysis, and also with the degree of coupling.
decreasing linearly from unity poleward of latitude 30° to zero equatorward of 15° latitude.

Keenan et al. (1986) describe a univariate 1° × 1° resolution screen-level temperature and dewpoint analysis scheme. These analyses are merged with the tropospheric analysis fields over land to provide a more detailed low-level temperature and moisture analysis. Over the oceans the 1000 hPa temperature analysis is regressed towards the climatological sea-surface temperature to prevent any drift in the analysis in these data-sparse areas.

The forecast model used to provide the guess fields is the Australian Region Primitive Equations (ARPE) model described by Leslie et al. (1985) and Mills and Leslie (1987), and was configured with the same horizontal grid spacing as the analysis system, but with 11 sigma levels (0.98, 0.95, 0.9, 0.85, 0.70, 0.50, 0.40, 0.30, 0.20, 0.10, 0.05). Because the analysis is performed on pressure surfaces and the prediction model is formulated in sigma coordinates, vertical interpolations are necessary at each interface in an analysis/forecast cycle. It is clearly desirable that in data-void areas, the prediction model fields on sigma surfaces should be unaltered by the vertical interpolation. This is achieved by interpolating increments (i.e., analysis minus first-guess differences on pressure surfaces), rather than full fields, from the analysis to the forecast coordinates, using cubic splines. The forecast fields on sigma surfaces are adjusted for changes in surface pressure during the analysis, and the interpolated increments are added on the new sigma surfaces. A horizontal filter (Phillips 1979) is applied to these fields to remove two-grid length noise, and the model is integrated to the next analysis time. The vertical mode initialization scheme of Bourke and McGregor (1983) is used at the commencement of the forecast. Lateral boundary conditions were provided for the forecasts from the archived NMC operational Southern Hemisphere forecasts (Bourke et al. 1982).

The output of this system is designed to be a set of analyses that fit the data according to the specified errors of each data type, which preserve forecast model generated detail in areas lacking other data, which also preserve the forecast model generated divergence fields unless the data indicates otherwise, and which are primarily used to initialize the NMC operational limited-area NWP model. Table 2 shows the root-mean-square fit of a selection of analysis fields to data from the differing observing platforms. The data are being fit by the analysis to the approximate order of the externally specified observational errors, as would be expected.

<table>
<thead>
<tr>
<th>Field</th>
<th>Data type</th>
<th>Rms difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>MSL (hPa)</td>
<td>SYNOP</td>
<td>0.74</td>
</tr>
<tr>
<td>MSLP (hPa)</td>
<td>Ship</td>
<td>1.24</td>
</tr>
<tr>
<td>850–700 hPa layer</td>
<td>TOVS</td>
<td>8.86</td>
</tr>
<tr>
<td>thickness (gpm)</td>
<td>TOVS</td>
<td>11.19</td>
</tr>
<tr>
<td>500–400 hPa layer</td>
<td>TOVS</td>
<td>9.80</td>
</tr>
<tr>
<td>thickness (gpm)</td>
<td>TOVS</td>
<td>13.18</td>
</tr>
<tr>
<td>300–250 hPa layer</td>
<td>TOVS</td>
<td>11.15</td>
</tr>
<tr>
<td>thickness (gpm)</td>
<td>TOVS</td>
<td>2.34/2.28</td>
</tr>
<tr>
<td>850 hPa zonal wind</td>
<td>RAWIN</td>
<td>2.62/2.66</td>
</tr>
<tr>
<td>meridional wind (m s⁻¹)</td>
<td>RAWIN</td>
<td>2.48/2.45</td>
</tr>
<tr>
<td>500 hPa zonal wind</td>
<td>SINGL</td>
<td>3.14/3.01</td>
</tr>
<tr>
<td>meridional wind (m s⁻¹)</td>
<td>SINGL</td>
<td>5.14/5.16</td>
</tr>
</tbody>
</table>

sections of this paper indicate that this is the case with the analyses used.

The analyses from the period from 1200 UTC 25 August to 0000 UTC 28 August 1985 will be used in the remainder of this paper, and date/time conventions used will be of the form 28/0000, where the first number is the day in August 1985, and the second is UTC time.

3. Synoptic setting

a. Mean sea level and cloud band evolution

The sequence of MSLP and 900 hPa isentrope analyses, and GMS infrared satellite imagery at 12-hour intervals from 25/1200 to 28/0000 are shown in Figs. 3 and 4. (The availability of both pressure and sigma-surface analysis fields enables the use of 900 hPa as a display level. It is chosen since it is just above all model topography over the Australian region.) At 25/1200 a deep low was south of Western Australia (WA), and a weakening trough was over Victoria (the remnants of a cutoff low, which had formed over WA some days earlier). The subtropical ridge extended across the continent between 25° and 30°S. A long, narrow frontal cloud band was associated with the deep Southern Ocean low, with an area of cold air to its west, and a broad area of stratiform cloud east of the cloud band.

As the surface low moved eastward, and then southeasterwards, the associated trough approached the southwest corner of WA. As it did so the trough sharpened and pressure fell over WA until by 27/0000, a marked pressure trough extended through WA to the northwest coast. During the final 24 hours of the period the trough moved eastwards across Australia at some 12–15 m s⁻¹, and was located over Victoria by 28/0000. The sharpening of the trough is shown both by the time sequence of maximum 900 hPa cyclonic rel-
Fig. 3. Mean sea level pressure analyses at 12-hour intervals from 1200 UTC 25 August 1985 to 0000 UTC 28 August 1985, with 900 hPa isentropes (dotted) overlaid.

ative vorticity at 30°S (Fig. 5), with the cyclonic vorticity increasing from $-9 \times 10^{-6}$ to $-69 \times 10^{-6}$ s$^{-1}$ (approximately 0.12/ to 0.95/) at 27/0000, and then slowly decreasing as the trough moves eastwards, and also by the sequence of 12-hour mean-sea-level isallobaric fields shown in Fig. 6. The 12-hour interval was chosen to reduce the effect of diurnal pressure variations. At 26/0000 there is a fall center over southwest WA, indicative of the sharpening trough, and another fall center well south of the continent associated with the deep Southern Ocean low. By 26/1200 the zone of pressure falls over WA has moved eastwards and
Fig. 4. GMS infrared satellite imagery at 12-hour intervals from 1200 UTC 25 August 1985 to 0000 UTC 28 August 1985.
extended northwards, with the suggestion of a secondary maximum near 25°S, 120°E. Thereafter the main center moves southeastwards. It should be noted that sallobaric gradient is much stronger on the western than the eastern side of the trough, and this point will be returned to later in this paper.

At 25/1200 there are two zones of enhanced thermal gradient at 900 hPa. The first is associated with the trough over southeastern Australia, and extends through central Australia on the northern flank of the small anticyclone centered near 28°S, 130°E, finally crossing the WA coastline on the northern side of the ridge extending from the Indian Ocean anticyclone. This zone is well marked by the 296 and 300 K isentropes in the first panel of Fig. 3. The second baroclinic zone is associated with the approaching southern trough and is roughly aligned with the 288 K isentrope.

The analyses used are only as good as the system (data, guess field, and analysis technique) that produced them. Thus, the low-level temperature gradients over the continent, which are strongly constrained by the surface observing network, should be quite reliable, allowing for the grid spacing used. Less reliance should be placed on the magnitude of the thermal gradient over the ocean, however, as the following factors all impact this. First, the only thermal data available over the ocean are the TOVS data, which have insufficient horizontal resolution (see Fig. 2) to unambiguously define the thermal gradient over the sea, although the forecast model does generate baroclinic zones in the fields used as first guesses in the analysis. Further, the lowest analysis level temperatures are constrained by the climatological sea-surface temperatures, which will reduce thermal gradients in the lowest levels over the ocean, although this is not wholly unrealistic, as can be seen in the time section of thermal structure across a front over ocean and land shown by Wilson and Ryan (1987). In spite of all these factors the position of the baroclinic zones, if not the intensity, is probably reliable over the ocean, and useful information, at least qualitatively, can be gleaned from these analyses.

At 26/0000 the baroclinic zone over the continent had not moved a great deal, but that associated with the approaching southern trough was nearing southwestern WA, having moved some 10 degrees eastward in the previous 12 hours. There is still a well-defined zone of weak temperature gradient over WA between these two baroclinic zones. By 26/1200 the two baro-

![Figure 5](http://journals.ametsoc.org/mwr/article-pdf/117/7/1402/4170689/1520-0493(1989)117_1402_doarcb_2_0_co_2.pdf)

**Fig. 5.** Time sequence of maximum 900 hPa cyclonic relative vorticity (s⁻¹ × 10⁶) on the axis of the trough at 30°S (full line) and maximum value of |∂θ| at 30°S (dotted).

![Figure 6](http://journals.ametsoc.org/mwr/article-pdf/117/7/1402/4170689/1520-0493(1989)117_1402_doarcb_2_0_co_2.pdf)

**Fig. 6.** Sallobar fields (hPa) over 12-hour intervals centered at 0000 and 1200 UTC 26 August and 0000 UTC 27 August 1985.
Clinic zones had merged, with the 288 K isentrope moving northeastwards and the 296 K isentrope moving southwestwards. The thermal gradient over WA further strengthened in the 12 hours to 27/0000 (note the position of the 288 and 304 K isentropes) and the baroclinic zone then moved eastwards across Australia in the following 24 hours. The changes in thermal gradient associated with the baroclinic zone can be seen in Fig. 5 where the maximum 900 hPa potential temperature gradient at 30°S at each 6-hour analysis time is plotted. These gradients were calculated using fourth-order differencing on the 250 km grid, and the values plotted in Fig. 5 interpolated by eye from the resulting contoured fields. The steep increase between 26/0000 and 26/1200 can be clearly seen.

As an example of how the passage of this baroclinic zone was seen at individual stations in WA, the temperature traces obtained from thermograph records at Perth, Meekatharra and Kalgoorlie (see Fig. 1), together with three-hourly 10 m wind observations are shown in Fig. 7. At Perth the initial temperature change was very sharp, some 4 K in only a few minutes, followed after about an hour by a further steady temperature fall of some 6 K in 3 hours, probably with some diurnal contribution. The sharp wind change with the first temperature change and the observed light showers are also shown. The front passage, as defined as the start of the rapid temperature change, was around 0400 local time at both Meekatharra and Kalgoorlie. Each showed a temperature increase, against the diurnal trend, in the hour before the front passed, and with a greater, but less rapid, cooling than at Perth with the initial cool change. Neither of these two latter stations reported rainfall, and, indeed, the reported falls were only over the far southwest of WA, with 2 mm or less being a typical rainfall report.

The satellite imagery associated with this chart sequence (Fig. 4) shows the narrow frontal cloud band moving steadily eastwards across the Indian Ocean, crossing the southwest corner of WA around 26/1200. Most of the continent was clear at this time, but some traces of cloud were just discernable in the southwest of the Northern Territory (NT). By 27/0000 this cloud area was more organized, and two small areas of middle level cloud were developing ahead of the frontal band near the coastline, while the northern part of the frontal band was beginning to weaken and dissipate. Rapid evolution of the cloud field took place in the next 24 hours. The frontal band thickened over the ocean south of SA, while the northern part of the old frontal cloud band had completely dissipated by 27/1200. The cloud areas ahead of the old frontal cloud band continued to develop, and between 27/1200 and 28/0000 formed a cloud band some 10° longitude wide over southeastern Australia. Comparing Figs. 3, 4 and 6 shows that the separation of the MSLP isolobes along the baroclinic zone between 26/1200 and 27/0000 followed the establishment of the two isallobaric fall centers at 26/1200. A small low center had formed near 36°S, 128°E by 27/0000, and this then moved south-eastward along the baroclinic zone during the next 24 hours, following the isallobaric fall center. This fall center/low center has a close association with the cloud area that developed near 30°S, 125°E ahead of the main frontal cloud band at 27/0000 and the subsequent thickening of the frontal cloud mass south of the continent at 27/1200 and 28/0000.

The first reports of rainfall were made over the SA gulls around 27/1800 and by 28/0000 rain or thunderstorm reports extended from north-central SA through to the western half of Victoria. Figure 8 shows the reports of rainfall and present weather at 28/0000, and a further 5–15 mm were reported over central Victoria in the following 6 hours.

**b. Low-level flow**

The 900 hPa vector wind fields are shown in Fig. 9 at 12-hour intervals from 25/1200 to 28/0000. Initially
the flow is fairly light over all but the southeast of the continent, although there is an elongated, but not intense, easterly flow maximum across WA associated with the east–west baroclinic zone (Fig. 3). As the surface trough moved into WA and amplified, and the thermal ridge ahead of the trough also amplified, there was an acceleration of the northeasterly flow on the eastward side of the thermal ridge, an acceleration of the northwesterly flow ahead of the trough, and an acceleration of the southerly flow to the west of the trough. By 27/0000 the northwesterlies ahead of the trough and the backing of the northeasterlies over central Australia have combined to form a low-level northerly jet ahead of the trough, with speeds exceeding 20 m s$^{-1}$, while a broad southerly low-level jet has formed over WA, with peak speeds up to 25 m s$^{-1}$. Note that the strengthening southerly flow can be associated with the strong isallobaric gradient west of the region of pressure falls (see Fig. 6).

c. Jet streak evolution and vertical motion

As it is often instructive to interpret vertical motion in the context of jet streak circulations (as in the recent reviews by Bluestein 1986; Keyser and Shapiro 1986; Uccellini and Kocin 1987), these two fields will be described in this section. Figure 10 shows the 300 hPa isotach and geopotential contour analyses. At 25/1200 the axis of the upper trough was located just west of the surface low near 100°E. An isotach maximum extended around the apex of the trough, linking in the north with the subtropical jet that extended across the continent. The strongest analyzed wind speeds at this time at 300 hPa were about 65 m s$^{-1}$ over central Queensland.

As the upper trough approached the WA coast, it also sharpened with the jet maximum west of the trough moving to its apex and increasing in strength to more than 60 m s$^{-1}$ by 26/0000. At this time a marked "split" in the jet structure was evident east of the trough, with a northwesterly stream extending southwards along the eastern flank of the trough and a westerly streak linking the isotach maximum at the apex of the trough with the subtropical jet. As the upper trough moved eastwards, the flow over Australia became more meridional; the jet streak moved from the apex to the eastern flank of the trough and strengthened to more than 70 m s$^{-1}$ over southern WA by 27/0000. With the increasing meridionality of the upper flow, the jet streak propagating around the upper trough became the dominant isotach feature over Australia, with only vestigial connection with the subtropical jet remaining. The upper trough–jet streak system then moved eastward between 27/0000 and 28/0000, with the jet streak remaining on the eastward flank of the trough near 30°S and slowly weakening from ~70 m s$^{-1}$ to ~55 m s$^{-1}$ during that 24 hours.

The vertical motion fields at 700 hPa are shown in Fig. 11. At 25/1200 the main centers of ascent south and southwest of WA are located near the equatorward entrance and poleward exit regions of the jet streak at 38°S, 110°E. By 26/0000 these areas of ascent have moved east and southeast respectively, and the northern center, now close to the southwest corner of WA, has increased in amplitude and is located between the poleward exit region of the jet at the apex of the trough and the equatorward entrance of the southern jet streak. Superficially, it may appear that this ascent maximum moves only very slowly eastwards in the following 12 hours to 26/1200, but the association with the jet streak shows that the ascent maximum between the two jet streaks moves from near 32°S, 115°E at 26/0000 to approximately 38°S, 128°E at 26/1200, and the center near 30°S, 116°E is actually a new ascent maximum associated with the circulations at the entrance region of the reorganizing northwesterly jet streak. Note also the strong subsidence region on the poleward entrance side of this jet streak. Thereafter this ascent maximum remains the dominant region of ascent over the continent and maintains its association with the entrance region of the jet, moving first north-eastwards and then southeastwards across Australia. By 27/1200 a further ascent maximum, near the poleward exit region of that jet, has become well established over the ocean south of the Great Australian Bight. The patterns at 28/0000 show a weakening of the jet maxima to below 55 m s$^{-1}$ and the formation of two streaks over the continent (see Fig. 10). It should be noted that the organization of the northwesterly jet over WA around 26/0000–26/1200 and the development of the major ascent maximum at its equatorward entrance region precede the sharpening of the surface trough (see Fig. 5) and the northward extension of the isallobaric fall center (Fig. 6). Allowing that the isallobaric fields are calculated over 12-hour intervals and that the omega fields are valid at analysis time, one can identify an association between the vertical
motion patterns and the isallobaric fields. It should be noted also that, while the preceding discussion has concentrated on the centers of maximum vertical motion, there is a broad general area of ascent associated with the baroclinic zone and that the two centers of maximum 700 hPa ascent at the entrance and exit re-
regions of the northwesterly jet streak are in this zone and are not separated by any region of subsidence.

4. Frontogenesis, ageostrophic flows and transverse circulations

In the previous section the evolution of the low-level pressure, wind and potential temperature fields were described, and vertical motion fields were described in the context of their associations with the jet streaks. So far no mention has been made of the source of moisture for the developing cloud band. Before we do this, however, it may be instructive to attempt to link together some of the processes which were described in the previous section, as it was shown that the thermal
**FIG. 11. Vertical motion (hPa h⁻¹) fields at 700 hPa at 12-hour intervals from 1200 UTC 25 August to 0000 UTC 28 August 1985.**
gradient increased over WA, pressures fell and that the low-level flow accelerated as and immediately after the northwesterly jet streak established itself over WA.

Petterssen (1940) introduced a frontogenesis function $F$ as the rate of change of potential temperature gradient following an air parcel. Following Miller (1948) this can be expressed on an isobaric surface as

$$F = \frac{d}{dt} \left| \nabla_p \theta \right| = \frac{1}{|\nabla_p \theta|} \left\{ - \frac{1}{2} \left[ \left( \frac{\partial \theta}{\partial x} \right)^2 + \left( \frac{\partial \theta}{\partial y} \right)^2 \right] \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) \right\} A$$

$$- \frac{1}{2} \left[ \left( \frac{\partial \theta}{\partial x} \right)^2 - \left( \frac{\partial \theta}{\partial y} \right)^2 \right] \left( \frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right) \}
$$

$$- \frac{\partial \theta}{\partial x} \frac{\partial \theta}{\partial y} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) B$$

$$+ \frac{\partial \theta}{\partial x} \frac{\partial \theta}{\partial x} \left( \frac{\partial v}{\partial y} + \frac{\partial \theta}{\partial y} \right) C$$

$$+ \frac{\partial \theta}{\partial x} \frac{\partial \theta}{\partial x} \left( \frac{dB}{dt} \right) + \frac{\partial \theta}{\partial y} \frac{\partial \theta}{\partial y} \left( \frac{dB}{dt} \right) . D \ (1)$$

The terms on the right-hand side of the equations are respectively the contributions to frontogenesis due to convergence ($A$), deformation ($B$), tilting ($C$), and diabatic forcing ($D$).

Fields of $F$ at 900 hPa were calculated at 12-hour intervals using Eq. (1), with the diabatic terms assumed zero; these are shown in Fig. 12. At 26/0000, the zone of frontogenesis has just crossed the WA coastline, and in the next 24 hours both intensifies and extends northwards as the main center moves north-eastwards and a second maximum forms near 41°S, 133°E. Terms $A$, $B$, and $C$ of Eq. (1) were evaluated at each analysis time, and fields of the contributions to $F$ from convergence (term $A$) and deformation (term $B$) are shown in Fig. 13 during the period of rapid synoptic evolution. The contribution due to convergence increases markedly over southwest WA between 26/0000 and 26/1200 and then by another 25% to 27/0000 as its maximum moves north-eastwards and is closely associated with the developing ascent maximum at the entrance region of the jet. The contribution to frontogenesis from deformation has its maxima oriented along the axis of the low-level pressure trough and has formed two major centers near 27°S, 114°E and 37°S, 122°E by 26/1200. Both centers move east-southeast in the following 12 hours to 27/0000. The northern center in particular can be associated with the shear zone to the east of the strong southerly flow to the rear of the baroclinic zone (see Fig. 9). It thus appears that two factors aided the intensification of frontogenesis over WA, although both were associated. The deformation contribution was forced by the increased shear associated with the accelerated southerly flow west of the trough, while the contribution due to convergence was associated with the vertical circulation at the entrance region of the jet streak. The maximum in the contribution due to convergence lay between the two maxima due to deformation thus producing an elongated frontogenetic zone.

It has been shown that there was a significant acceleration of both southerly flow to the rear of the front and of northerly flow ahead of the front between 26/1200 and 27/0000, and that the locations of vertical motion maxima suggest that they were associated with the transverse ageostrophic circulations which would be expected at the entrance and exit regions of jet streaks (for example, see Fig. 23 of Keyser and Shapiro 1986). Let us first look at the 900 hPa ageostrophic winds at 27/0000, chosen as typical of the period of reorganization, then look at selected cross sections to see what features of the low-level ageostrophic flow can be associated with transverse circulations.

Figure 14 shows the 900 hPa ageostrophic wind, calculated as the difference between the analyzed wind and the geostrophic wind at 27/0000 and the 900 hPa isallobaric wind over a 12-hour period centered at 27/0000. There is a strong convergent zone in the ageostrophic wind field over WA, along the line of the trough at 900 hPa, with greater than 10 m s$^{-1}$ north-easterly ageostrophic wind covering a large area east of the trough over central Australia, and with 5–8 m s$^{-1}$ south-westerly behind the trough. The isallobaric south-westerly winds are very strong (greater than 15 m s$^{-1}$), as would be expected from the strong isallobaric gradients to the west of the trough shown in Fig. 6, but are opposed by the inertial–advective component (not shown). East of the trough there is a clearly seen maximum in the north-eastern isallobaric wind, which is weakly reinforced by the inertial–advective wind, and the area of strongest isallobaric northeasterlies closely matches the area of maximum north-easterly ageostrophic flow. Both maximum isallobaric northeasterlies and south-westerlies converge in the area of the 700 hPa ascent maximum and thus appear to be associated with the circulations at the jet entrance.

Cross sections of ageostrophic wind parallel to the section and vertical velocity along the line A-B in Fig. 14 are shown in Fig. 15, this line chosen to be perpendicular to the 300 hPa geopotential contours at the jet entrance and to intersect the maximum area of rising motion shown in Fig. 11. The cross sections of potential temperature and total wind speed are also shown to indicate the position of the front and jet stream axis. A strong core of rising motion is seen, sloping southwards (left) with height, while the low-level ageostrophic flow shows strong convergence from north (left) and south (right). The upper branch of the circulation indicates a strong southerly transverse ageostrophic flow, with one maximum displaced just south of the jet core, although the transverse ageostrophic wind on this section is all directed south at jet-core
Fig. 12. Fields of frontogenesis function \((K \, m^{-1} \, s^{-1} \times 10^{-11})\) at 900 hPa at 12-hour intervals from 1200 UTC 25 to 0000 UTC 28 August 1985.
Fig. 13. Fields of contribution to frontogenesis at 900 hPa from deformation (right) and convergence (left) at 0000 UTC and 1200 UTC 26 August and 0000 UTC 27 August 1985. (Units K m$^{-1}$ s$^{-1}$ $\times$ 10$^{-11}$).
strong convergence in the low levels and the ascending air is clearly seen, with the upper branch of the direct circulation perhaps more clearly defined than that of the indirect circulation. Figure 15, which has sections normal to the upper level geopotential contours and to the lower level isentropes respectively, shows strong evidence for a direct circulation cell on the poleward side of the jet/baroclinic zone. There is a less well-defined indirect circulation, and the low-level ageostrophic flow from the left of each section is of a magnitude similar to that of the isallobaric wind. (Note that these sections do not intersect the strongest area of subsidence.)

While this analysis is not in sufficient depth to determine unambiguously precise roles of jet-associated and front-associated circulations, it does appear that the principal cause of the accelerated low-level flow was the pressure falls associated with the jet-streak circulations and that the frontogenesis over WA was largely a response to the divergence patterns and increased deformation associated with these jet-streak circulations.

5. Moisture availability

As an indicator of the availability of low-level moisture for cloud formation, the 900 hPa mixing ratio fields at 12-hour intervals from 26/0000 to 27/1200 are shown in Fig. 16. A broad moist tongue associated with the surface trough lies over WA, and a large area of very dry air lies over the central and southeastern parts of Australia. In the 24 hours to 27/0000 the moist tongue narrows and moves slowly eastwards, the area of dry air maintains its position but the moisture gradient to its north moves slowly southwards. In the following 12 hours, under the influence of the strengthening low-level northerly flow, there is a pronounced southward advection of high mixing ratio air over central WA, with the 8 g kg⁻¹ isopleth extending from around 21°S to 29°S. The axis of the moist tongue is in close proximity to the band of ascent at 700 hPa, and thus the isobaric coordinate system is not perhaps the best way of observing moisture field evolution. Benjamin (1987), amongst many others, has indicated the advantages of isentropic coordinates in this regard. Figure 17 shows the pressure, wind vectors, and relative humidity on the 305 K isentropic surface at 27/0000 and 27/1200. The band of high relative humidity air ascending in the northwesterly flow ahead of the trough, with its maximum near the south coast of WA at 27/0000, is apparently associated with the thickening cloud there. The thickening cloud south of SA at 27/1200 is also well represented on this isentropic surface, and the origins of the moist band in the continental boundary layer are clearly seen.

6. Trajectory analysis

The isentropic charts shown in Fig. 17 do not approximate parcel trajectories, as they are not relative
with a time step of 30 min. Back trajectories were calculated to 36 hours before 28/0000 using the 6-hour sigma-surface analyses, with the velocity components interpolated in space using a bidirectional Bessel function interpolator, and interpolated linearly in time between successive analyses. Endpoints were selected at 0.5 grid unit intervals at 700 hPa (i.e., approximately each 125 km) over a rectangle encompassing the cloud band, shown in Fig. 18. Each trajectory was examined, and an arbitrary classification according to type was assigned. These types are listed in Table 3, with a description of their characteristics. Figure 18 shows the trajectory classification for each trajectory endpoint, while Fig. 19 shows examples of each trajectory type,
FIG. 16. Mixing ratio (g kg$^{-1}$) at 900 hPa at 12-hour intervals from 0000 UTC 26 August 1985 to 1200 UTC 27 August 1985.

with the pressure (hPa) of the parcel at 6-hour intervals. Figure 20 shows isopleths of 18-hour ascent (hPa) for the ascending parcels.

Kuo et al. (1985) performed sensitivity studies to study the errors in trajectory computation associated with temporal and spatial resolution of the data used to generate the trajectories. Insofar as their results can be extrapolated to the current data sets, errors due to any inadequacies of the 250 km, 6-hour resolution of the analyses should be less than 200 km and 20 hPa after 24 hours, and thus are insignificant to the conclusions drawn in this section of the paper.

Type E trajectories are associated with the ridge area ahead of the main trough, type C trajectories show descent to the rear of the front and mark the subsiding air. Trajectory types A, F and G appear to be associated with the "old" cloud band, and maintain a close, although naturally evolutionary, association with the band throughout its life. Class B trajectories originate in the boundary layer over Central Australia, and experience most of their ascent during the second 18 hours. It can be seen from Figs. 19 and 20 that these trajectories show the greatest rates of ascent and also (not surprisingly) have their endpoints in the middle of the main cloud mass. The class B trajectories are those which bear the closest similarity to the "warm conveyor belt" flows described by Harrold (1973), and it must be noted that these flows originate in the boundary layer over the northern interior of the continent, where the air mass has been well mixed by boundary layer processes. Differing from the A, F and G parcels, though, these trajectories are in clear air until the last 12 hours of their ascent, and appear to provide the source air (and moisture) for the thickening cloud mass. Figure 18 shows the envelope of B trajectory endpoints at 28/0000 and also the envelope of those air parcel positions 18 hours earlier at 27/0600, i.e., before the bulk of their ascent occurred. The early flow from the northeast of these type B trajectories and the position of the envelope of their positions at 27/
Fig. 17. Pressure and wind vectors (left) and relative humidity (right) on the 305 K isentropic surface at 0000 UTC and 1200 UTC 27 August 1985.

Fig. 18. Classification of trajectories ending at 700 hPa at the location of letters. Refer to Table 3 for the key.

0600 show that they are associated with the enhanced northeasterly flow, which was shown in the previous section to be largely isalohypsic, and then ascended in the northwesterly flow shown in Fig. 17.

The other ascending trajectories, A, F and G, are confined to a narrow band near the rear of the cloud.

### Table 3. List of trajectory classifications.

<table>
<thead>
<tr>
<th>Type</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Long, steady, straight ascent from WNW. Ascent possibly greater in first 18 hours.</td>
</tr>
<tr>
<td>B</td>
<td>Rapid ascent in last 18 hours. Flow curving from northeast to northwest.</td>
</tr>
<tr>
<td>C</td>
<td>Curving descent, origin far to southwest.</td>
</tr>
<tr>
<td>D</td>
<td>Straight flow from SW. Modest ascent or descent.</td>
</tr>
<tr>
<td>E</td>
<td>Straight slow westerly flow.</td>
</tr>
<tr>
<td>F</td>
<td>Ascent from west, followed by descent.</td>
</tr>
<tr>
<td>G</td>
<td>Curving ascent from the southwest.</td>
</tr>
</tbody>
</table>
band, and during the early part of their trajectory are associated with the low-level trough, although they tend to lag the (low-level) trough towards the end of the trajectory period, when they are nearing 700 hPa.

7. Conclusion

The synoptic evolution of the development and passage of a baroclinic zone across Australia has been described, with emphasis on an attempt to understand the mechanisms that lead to rapid cloud band development over southeastern Australia. Surface pressures fell over WA as a jet streak propagated around the apex of an upper trough, with a sharpening of the surface pressure trough as it moved from the ocean into WA. With the establishment of the northwesterly jet streak over WA, ascent maxima at both entrance and exit regions of the jet became well defined with the northern vertical circulation leading to low-level frontogenesis, both directly from the convergence term in Petterssen's frontogenesis equation and also by the increased deformation caused by the strengthened isallobarysic wind.

The moisture isopleths, initially approximately eastward across much of the continent, were advected southward by the accelerating northeasterly and northwesterly flow ahead of the trough. Back-trajectory analysis shows that most of the cloud, which developed over the continent and had its origin in the continental boundary layer over central Australia, was initially accelerated southwestward with the strengthening northeasterly flow and then lifted into the ascending northwesterly flow ahead of the trough. Thus these ascending trajectories may be regarded as a type of Harrold's (1973) "warm conveyor belt," with their characteristic being an origin in the boundary layer over central Australia, and rapid ascent over a period of 12–18 hours prior to cloud band formation.

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REFERENCES


