Aust. Met. Mag. 51 (2002) 203-221

A case of coastal interaction with a cool change

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(Manuscript submitted February 2002; revised July 2002)

A surging cool change moved along the Victorian coastline on the afternoon of 21 January 1997. A mesoscale numerical prediction of this event produced an excellent simulation of the evolution of this front, and is used to diagnose processes leading to the intensification and movement of the change. It is shown that the development of an internal boundary layer in hot, dry offshore flow can mask the thermal structure of the change at the lowest levels over the ocean, and can also provide a coastal thermal (density) gradient which, with the reversal of the coastal pressure gradient can lead to the surging of a front along the coast. Differing wind structures before and after the change are seen in the inland and coastal portions of the change, and the implications of these for operational forecasting are discussed.

Introduction

It has long been known that summertime fronts in southeastern Australia show pronounced diurnal changes in intensity, movement and times of arrival. As long ago as the 1940s this behaviour was linked to the coastal thermal and frictional gradient (Cassidy 1945; Loewe 1945). A series of field studies in the 1950s (Berson et al. 1957, 1959; Clarke 1961) demonstrated diurnal variations in speed and arrival time of southeastern Australia cool changes, and attributed some of these to the variations in land-sea heating contrast during the diurnal cycle.

Following the inception of meteorological satellite imagery it was recognised that fronts in southeastern Australia, particularly dry fronts during late spring and summer, can appear to rapidly accelerate across the coastline during the day, or 'surge' along the southern coastline of Australia. Other fronts can appear to have multiple change lines (e.g. Jasper and Taylor 1977). The Cold Fronts Research Program (Ryan et al. 1985), hereafter CFRP, was designed to address some of these issues. The CFRP observational programs were based in the late springtime period, and so tended to sample 'moist' changes, with multiple 'change lines' associated with convection. In addition, a series of studies addressing the issue of accelerating dry fronts was stimulated by the CFRP (e.g. Hanstrum et al. 1990a,b; Physick 1988; Reeder 1986; Garratt 1986,1988; Garratt and Physick 1986,1987; Garratt et al. 1989).

Hanstrum et al. (1990a,b) proposed the model of frontogenesis in the pre-frontal trough which develops over land ahead of the hyperbolic deformation zone between two anticyclones. In this model, they proposed an intensifying thermal gradient between pre-frontal oceanic west/northwesterly flow and hot continental northerly flow. Once the thermal gradient was established, quasi-geostrophic arguments were used to propose a positive feedback process of frontogenesis. The issue of what processes led to the formation of the trough was neglected, as were diabatic processes.

Physick (1988) used an idealised three-dimensional model of a front approaching a coastline to demonstrate that the relative phase of the sea-breeze circula-

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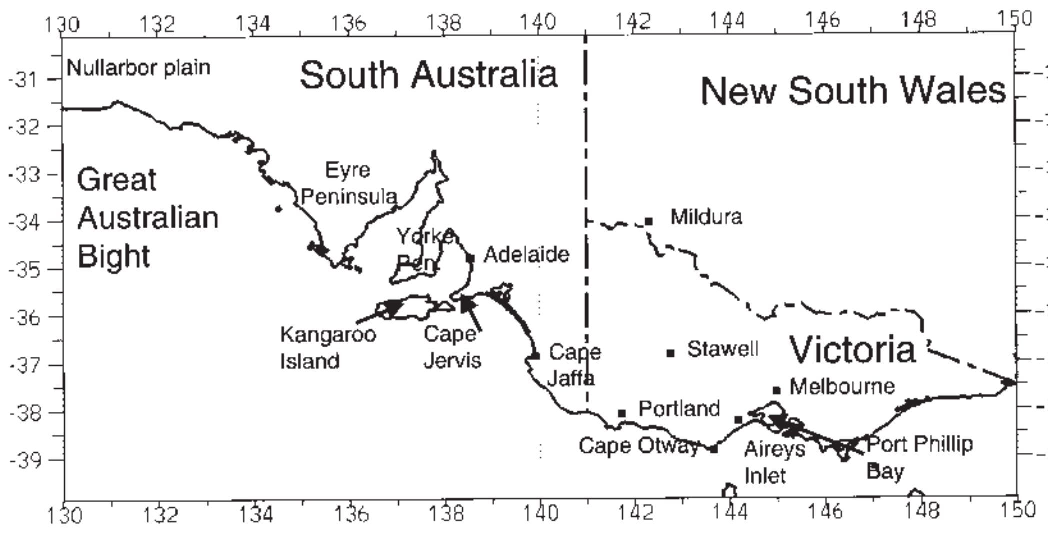
tion at the coast and the synoptic-scale front was important in determining whether the front would 'accelerate' as it approached the coast. He hypothesised that as a synoptic-scale front approached the coast during the daytime heating period, then the ascending branch of the cross-frontal circulation and the ascending branch of the sea-breeze circulation could reinforce each other to produce a stronger cross-frontal circulation and thermal gradient. Once this front crossed the coast, moving into the well-mixed overland daytime boundary layer, the cross-frontal scale of the frontal circulation would decrease and its intensity increase. This modified front would then become the cool change experienced through southern Australia. Reeder (1986) demonstrated this frontal contraction in his idealised two-dimensional modelling studies of a front moving into a well-mixed boundary layer. Common to both these studies is the importance to the coastal interaction process of the phase of the diurnal heating cycle as the Southern Ocean front approaches the coast. In each case the reduced static stability over the land during the daytime heating cycle leads to frontal contraction and an intensified cross-frontal vertical circulation, as might be expected from a qualitative analysis of the static stability term in the semi-geostrophic form of the Sawyer-Eliassen equation (Bluestein 1986, pp. 188-92).

Garratt (1986, 1988), Garratt and Physick (1986, 1987), and Garratt et al. (1989) addressed the issue of surging dry fronts along the southern coastline of Australia. They demonstrated that these fronts had some of the characteristics of gravity currents. Further, they argued that for particular offshore flow configurations, discontinuities in coastline orientation could generate offshore gradients in the depth of the maritime internal boundary layer (IBL). (Garratt (1990) presents a review of the processes leading to the development of internal boundary layers.) This density gradient, differential friction between land and sea, and the frontolytic effects of post-frontal diabatic heating over land could all contribute to the more rapid movement of a front over the sea relative to the land, with the portion over the sea developing the characteristics of a gravity-current. Other authors, (Baines 1980; McInnes and McBride 1993) showed that topographic blocking had an impact on the movement of cold fronts along a coastline.

Most of these studies were based on idealised models, or narrow samples, and detailed studies of the development and evolution of these surging fronts along the southern Victorian coastline have not followed.

On the evening of 21 January 1997 a shallow, dry cool change passed through Melbourne (see Fig. 1 for all place names used in this paper). This cool change had exhibited a 'surging' behaviour along the Victorian coast, particularly after it passed Cape Otway and reached Melbourne as a south-southeasterly change. An after-the-event simulation using a high resolution mesoscale Numerical Weather Prediction (NWP) model produced a very accurate representation of the event, and resolved rapid frontogenesis over southeastern South Australia during the morning, and the development of a dramatic southwesterly surge along the eastern coast of the Otway Peninsula during the afternoon.

Fig. 1 Locality diagram.



In this paper the model simulation will be presented, and validated against analyses of the surface data over southeastern Australia. The issue of the identification of a front in NWP model data will be discussed, as this is an important issue in identifying features in the following diagnoses. The processes leading to the frontogenesis over southeastern South Australia (SA) and the processes leading to the surge along the Otway coast will then be described using the high temporal and spatial resolution available in NWP datasets to identify processes unresolvable in the available observations. The processes so identified will be related to the various conceptual models reviewed above. Finally, the clearly different three-dimensional structures of the overland and coastal parts of the cool change will be described and the implications for forecasting will be discussed.

Identification of a front in NWP output

Before identifying a front in any dataset, it is necessary to define what is meant by a front. It has been the practice in Australia for synoptic-scale frontal analysis to be based on the principles enunciated by Guymer (1977). In this the purpose of the front is 'to outline on the surface analysis schematically a diagnosis of the mid-tropospheric temperature gradient wherever the data indicates that this gradient is sufficiently strong to create the PFJ (Polar Front jet) maximum, and where the presence of this jet is being shown by a surface discontinuity'. This practice fits well with satellite imagery interpretation over the ocean, but this definition would not show the shallow cool change of southern Australia as a front, although such a change may be the dominant forecast feature for the day.

At the time of the CFRP there was considerable discussion as to how to identify a front in NWP model output (e.g. in McInnes et al. 1994). Some of these difficulties resulted from the relatively coarse horizontal resolution used in the numerical models of the day. McInnes et al. (1994) recommended that the field of low-level relative vorticity be used to identify the front as this resolved the difficulty of the land-sea temperature gradient masking the frontal temperature gradient. Indeed, the necessary cyclonic curvature of the streamlines at the front makes this a reasonable choice. However, cyclonic vorticity alone is not a sufficient condition to identify a front.

The high horizontal resolution of today's mesoscale NWP models now allows the identification of the position of the cool change directly from the model's surface winds and low-level temperature gradients. Indeed, rather than avoiding the issue of the interaction of the cold-frontal thermal gradient with the coastal temperature gradient, the interaction between these two thermal gradients can be resolved. The Hewson (1998) definition of a front as 'the warm-air boundary of a zone of enhanced thermal gradient', subject to the thermal gradient being of sufficient strength places the 'front' at the point where significant cooling begins. This is also where the NWP output places the abrupt wind change, as will be shown later in this paper. It does not necessarily place the change on the axis of lowest pressure, or of maximum cyclonic relative vorticity, although this is usually the case. By its definition, it does not place the front on the zone of greatest temperature gradient. It has the great benefit of providing a simple, precise definition which lends itself to NWP model output, and which also represents the onset of cooling and marks the initial backing of the wind - the onset of the cool change.

Synoptic setting and model validation

The Limited Area Prediction System (LAPS, Puri et al. 1998) MSLP analyses at 12-hour intervals through the 24 hours to 1100 UTC 21 January are shown in Fig. 2. At 1100 UTC 20 January a Southern Ocean low with a pronounced tilt westwards with decreasing latitude was located over the waters south of the Great Australian Bight, and with a trough of low pressure over Eyre Peninsula extending northwestwards into the Nullarbor Plain. During the 24-hour period the Southern Ocean low moved slowly eastwards, while the trough ahead of it moved eastwards, through SA and into central Victoria, with considerable diurnal modulation – the amplitude of the trough weakened overnight and strengthened again during the daylight hours. The change that affected Victoria was associated with this trough passage. In the mid-troposphere, a deep trough was located to the southwest of the Southern Ocean low (Fig. 3) with a northwesterly oriented cloudband (Fig. 4) ahead of the trough. Both upper trough and cloudband moved eastwards during the 24 hours, with the eastern edge of the cloudband just reaching central Victoria by 1100 UTC 21 January. Both the upper and surface lows become almost fully cut-off near 40°S by 1100 UTC 21 January, with the southern portions of these troughs moving eastwards south of 45°S.

The model used is the LAPS model described by Puri et al. (1998) with the inclusion of a new planetary boundary-layer scheme, following Viterbo and Beljaars (1995). This model was configured for these forecasts with 0.10° grid spacing and 29 vertical lev-

Fig. 2 **LAPS mean-sea-level pressure analyses at 1100, 2300 and 1100 UTC 20, 21 January. Contour interval 2 hPa.**

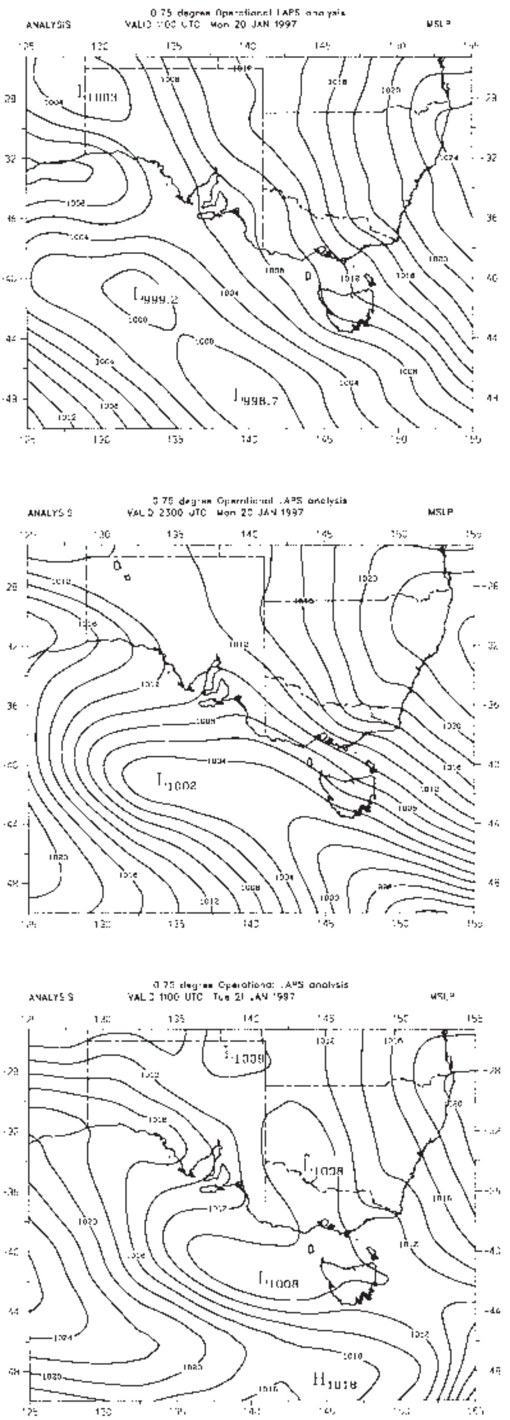


Fig. 3 **LAPS 300 hPa height/wind analyses at 1100, 2300 and 1100 UTC 20, 21 January. Height contours at 60 m intervals, wind speeds shaded at 10 m s^{-1} intervals.**

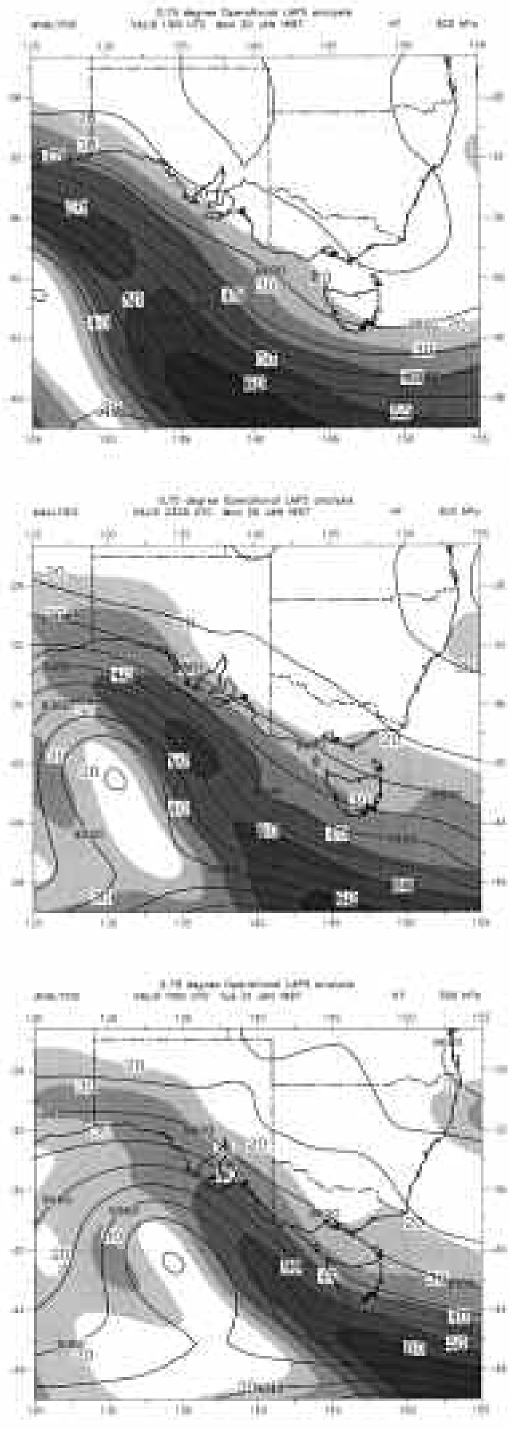
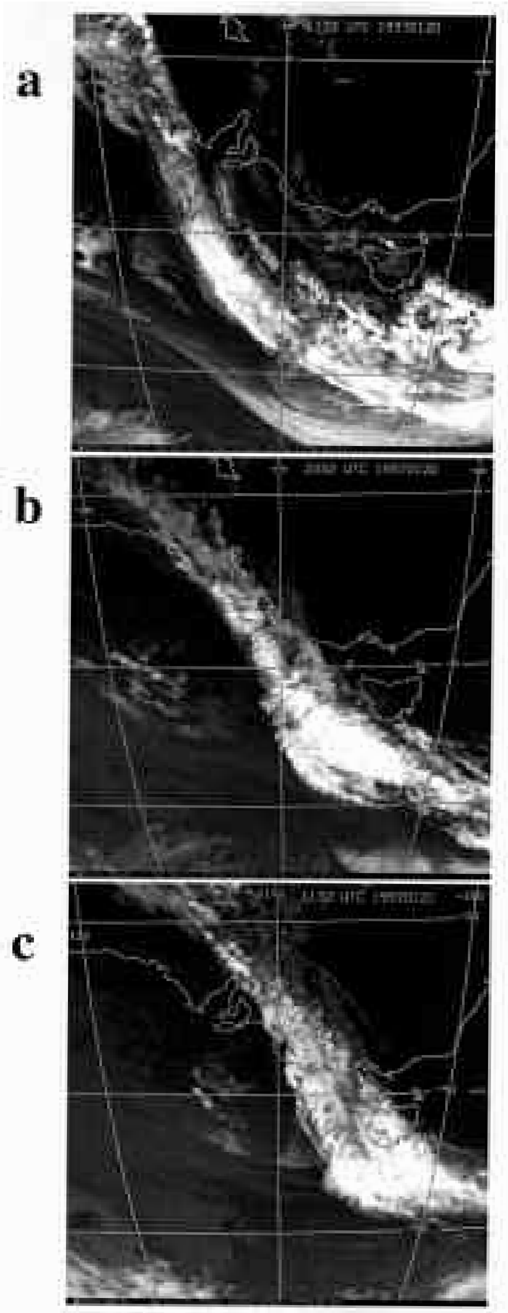


Fig. 4 GMS-5 IR satellite imagery at (a) 1132 UTC 20 January, (b) 2332 UTC 20 January and (c) 1132 UTC 21 January 1997.



els, and covered a domain of 30° longitude by 25° latitude centred over Victoria. This horizontal grid spacing is just a little less than that used since late 1999 in the operational mesoscale version of LAPS (0.125°), and has the same vertical level disposition, with the lowest level at approximately 10 m and 9 levels in the lowest 100 hPa of the atmosphere. Initial state and boundary conditions came from the operational LAPS forecast, which at that time had a horizontal grid spacing of 0.75° and 19 vertical levels. The model was initialised at 1100 UTC 20 January 1997, and run for 24 hours, with fields being output every hour to better resolve the interactions between the front and the land-sea interface.

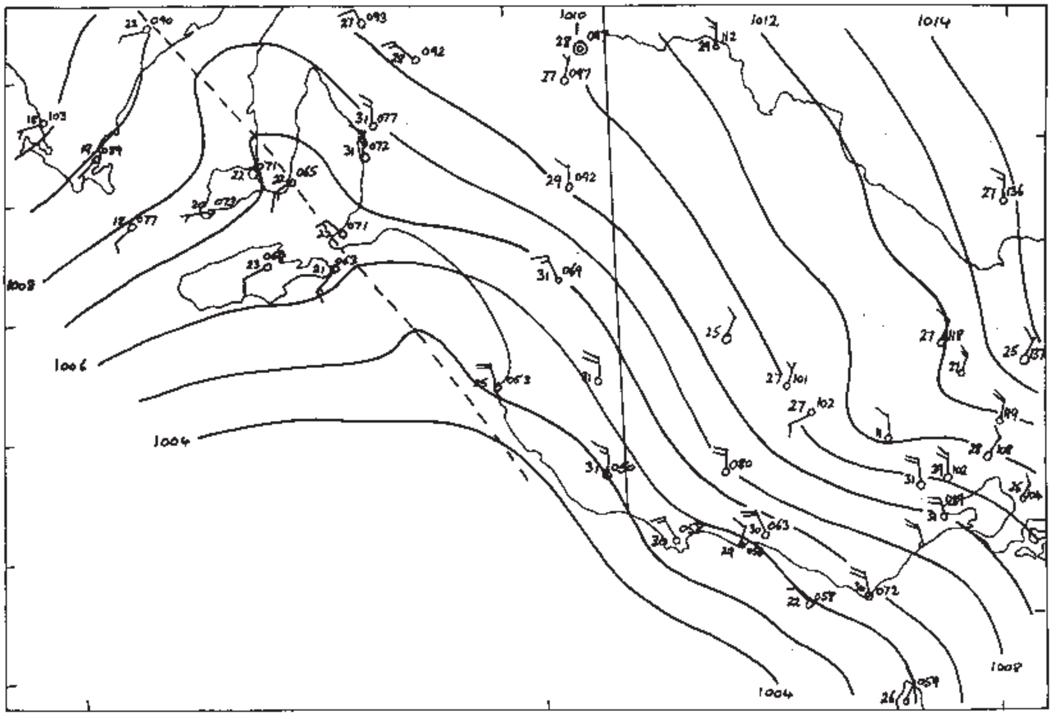
The forecast model was validated against subjective mesoscale re-analyses at 1900, 2200, 0100, 0400, and 0700 UTC on 20-21 January 1997. These analyses were based on the data used in the operational Victorian Regional Forecast Centre (RFC) analyses and some supplementary AWS data. They were analysed with the model forecasts as a guide to the pressure pattern, while still closely fitting the values of the pressure observations. While this might be seen as a lack of independence in the verification, it might also be seen as a paradigm for the use of mesoscale NWP model guidance in operational subjective mesoscale analysis. The rationale behind this argument is that the mesoscale model can, as will be shown in this paper, resolve processes that are not well sampled by the observational network. Thus the use of the model as a basis for the mesoscale analysis can assist the analyst in interpreting the observations. It must be noted, though, that the analysis must fit the observations unless there is good reason for not doing so.

At 1900 UTC 20 January 1997, the subjective reanalysis (Fig. 5(a)) shows a trough just approaching Cape Jervis, and extending through Yorke Peninsula. There is no evidence to locate this trough off the southeast coast of SA. The model forecast (Fig. 5(b)) shows this change marked by a pressure trough and the convergence of northerly and northwesterly wind vectors to be just a little west of the subjective placement between Kangaroo Island and Cape Jervis. This pressure trough and wind-change line can also be seen further south, just west of Cape Jaffa in the model forecast (Fig. 5(b)) but without any ship observations this cannot be verified. The model also indicates a second trough around 1.5° longitude west of the first, and this also marks a shift in wind direction from northwest to west. The forecast 10 m potential temperature field (Fig. 5(c)) shows the most significant thermal gradient to be located along the coast, with no definite thermal gradient associated with either of the pressure troughs or wind change lines.

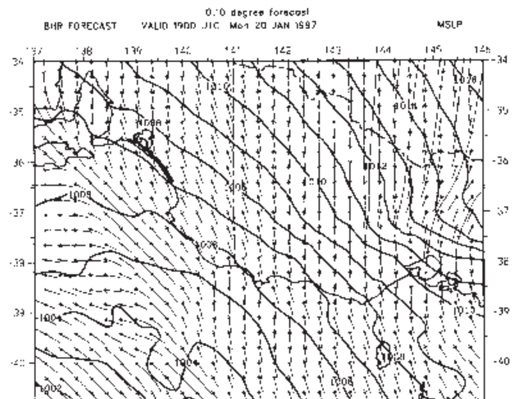
At 2200 UTC 20 January, the subjective re-analysis (Fig. 6(a)) shows the first pressure trough to have moved through Adelaide, and a wind change to have crossed the coast of southeastern SA, and to have passed Portland in western Victoria. The model (Fig. 6(b)) has good change position through the southeast of SA, and is just a little slow in moving it through

Fig. 5 (a) Subjective mesoscale mean sea-level pressure analysis at 1900 UTC 20 January 1997. Data plots have their usual meteorological meaning, contour interval 1 hPa. (b) Eight-hour mesoscale NWP model forecast of 10 m streamlines and mean sea-level pressure (contour interval 1 hPa) valid 1900 UTC 20 January 1997. (c) Eight-hour mesoscale NWP model forecast of 10 m streamlines and potential temperature (contour interval 2 K) valid 1900 UTC 20 January 1997.

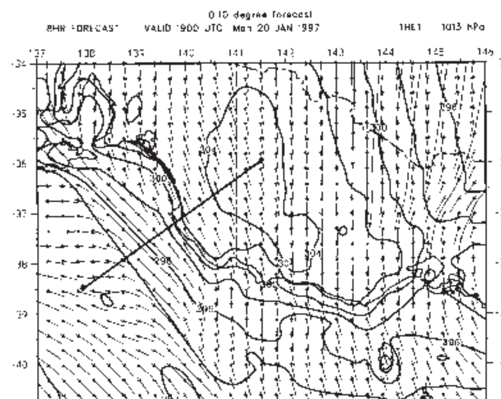
(a)



(b)



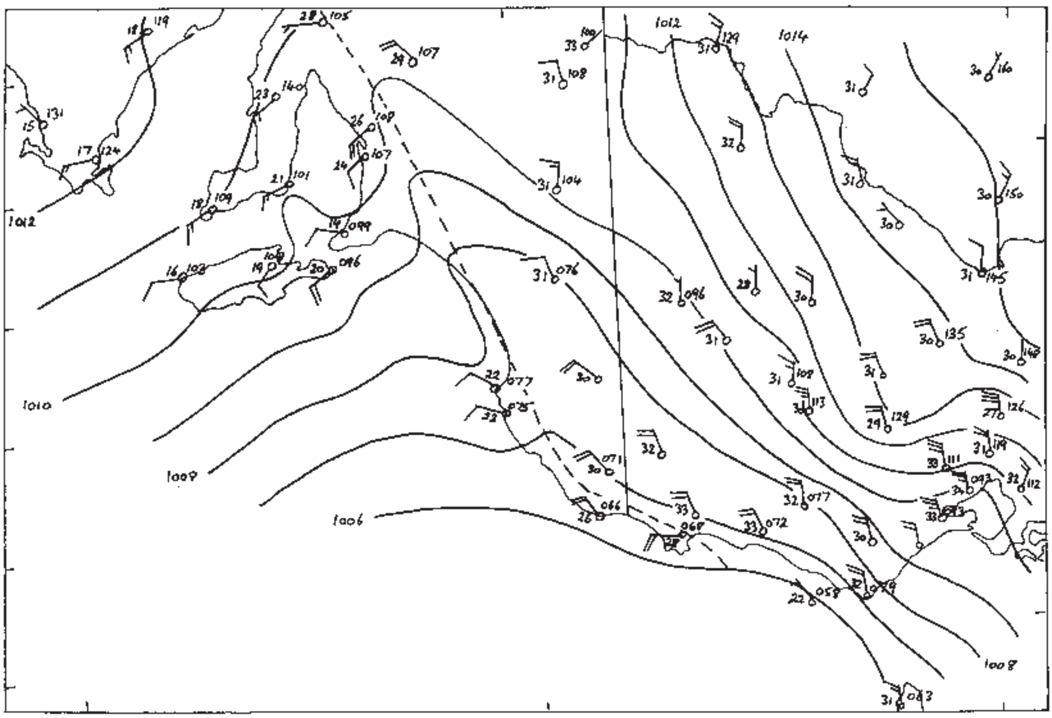
(c)



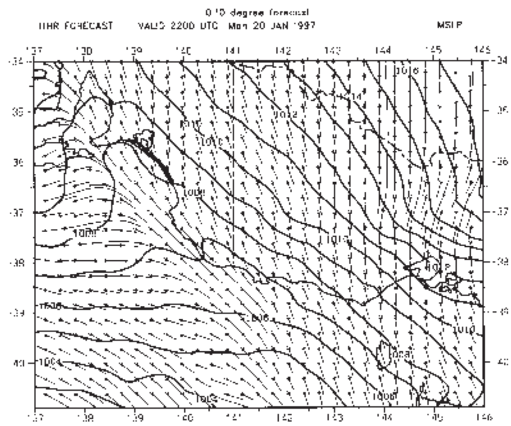
Portland. The model does not show a wind change to have moved through Adelaide as indicated by the observations, although there is a hint of a developing pressure trough just to the east of Adelaide. The model shows its major wind change to be associated with the western pressure trough that is located between Kangaroo Island and Cape Jervis in Fig. 6(b), some hint of which can be seen in the analysis (Fig. 6(a)). The thermal pattern (Fig. 6(c)) shows an increased coastal gradient, reflecting the commencement of solar heating during the preceding three hours, and with the 292K isentrope moving closer to the coast. Through the far southeast of SA, from Cape Jaffa to the Victorian border, the thermal gradi-

Fig. 6 As Fig. 5 for 2200 UTC 20 January 1997.

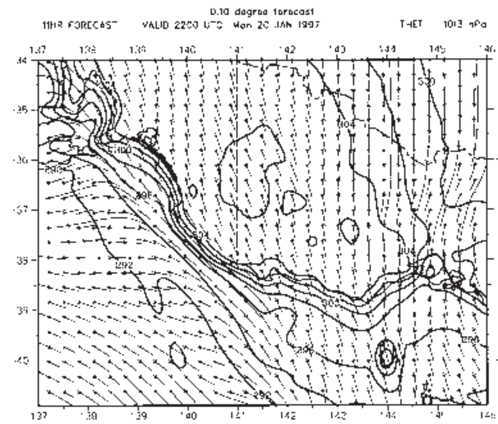
(a)



(b)



(c)

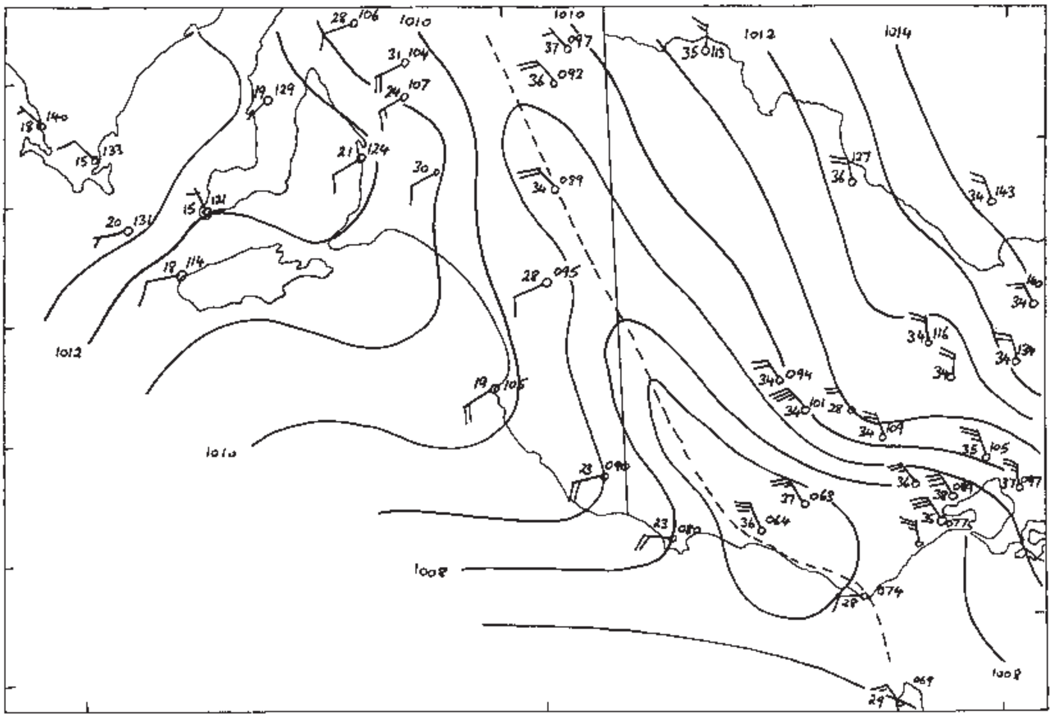


ent, and particularly the warm edge of the thermal gradient, has moved onshore. This is the only point at this time where there is a clear relation between the warm edge of a baroclinic zone and a wind change/pressure trough.

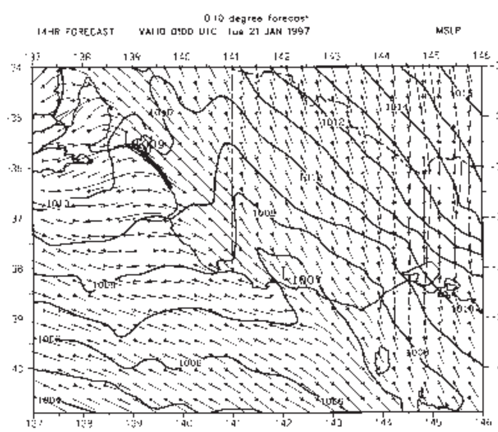
At 0100 UTC both the forecast (Fig. 7(b)) and the analysis (Fig. 7(a)) show a low-pressure centre forming over western Victoria, with a trough extending northwestwards. While the forecast winds are oriented northwesterly rather than south of west as indicated to the west of the trough by the observations, the simulated position and orientation of the low and trough are in excellent agreement with the analysis. The modelled thermal structure (Fig. 7(c)) has

Fig. 7 As Fig. 5 for 0100 UTC 21 January 1997.

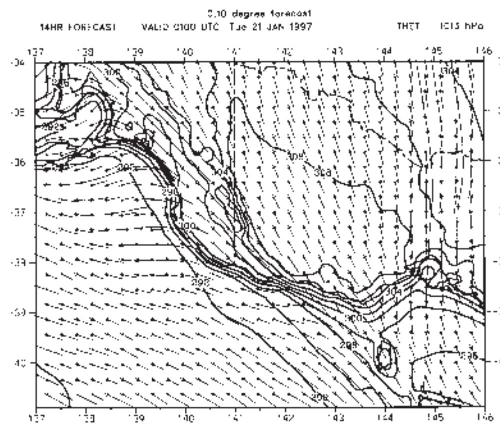
(a)



(b)



(c)

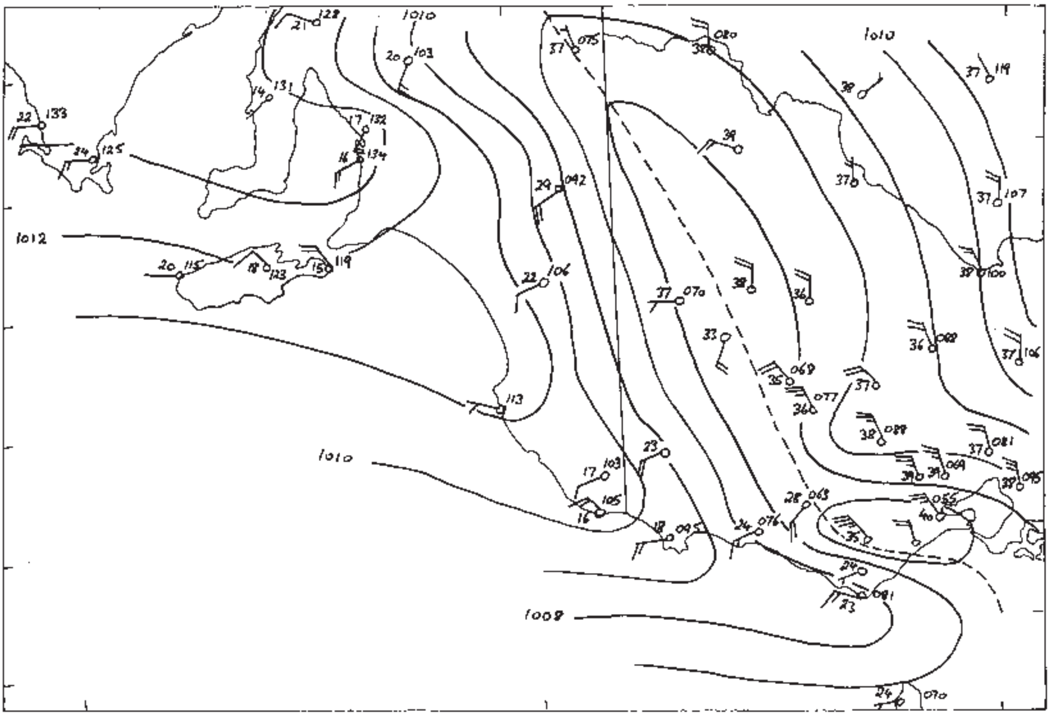


evolved dramatically in the three hours up to 0100 UTC. A marked thermal gradient has developed through western Victoria and southeast SA, with the wind change developing at its leading edge. While this is part of what would generally be termed a pre-frontal trough, for the remainder of this paper it will be termed the 'inland front', since it is being identified from the thermal and the wind fields, rather than by the pressure field. To the west of the surface trough, a very strong coastal thermal gradient exists along the western Victorian coast, and winds have backed from northwesterly just offshore. There is still a modelled temperature gradient along the southeast SA coast, and this is accompanied by broad cyclonic curvature of the wind field.

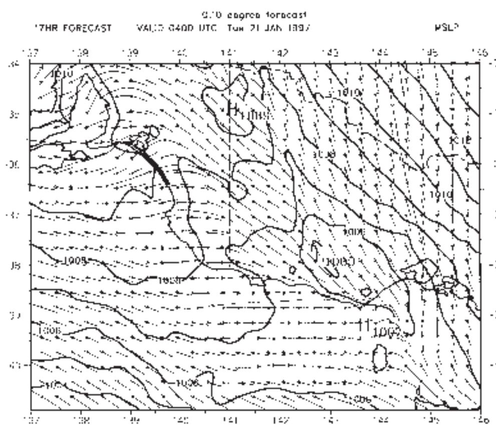
At 0400 UTC the verifying analysis and the model

Fig. 8 As Fig. 5 for 0400 UTC 21 January 1997.

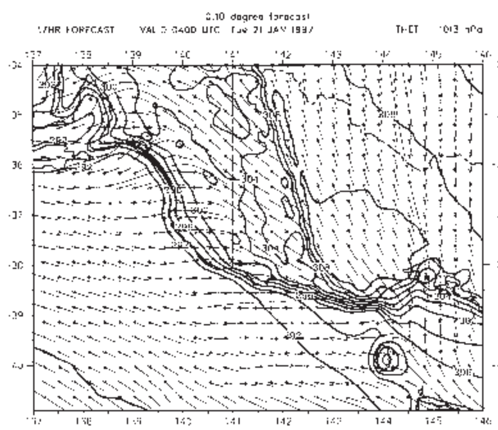
(a)



(b)



(c)

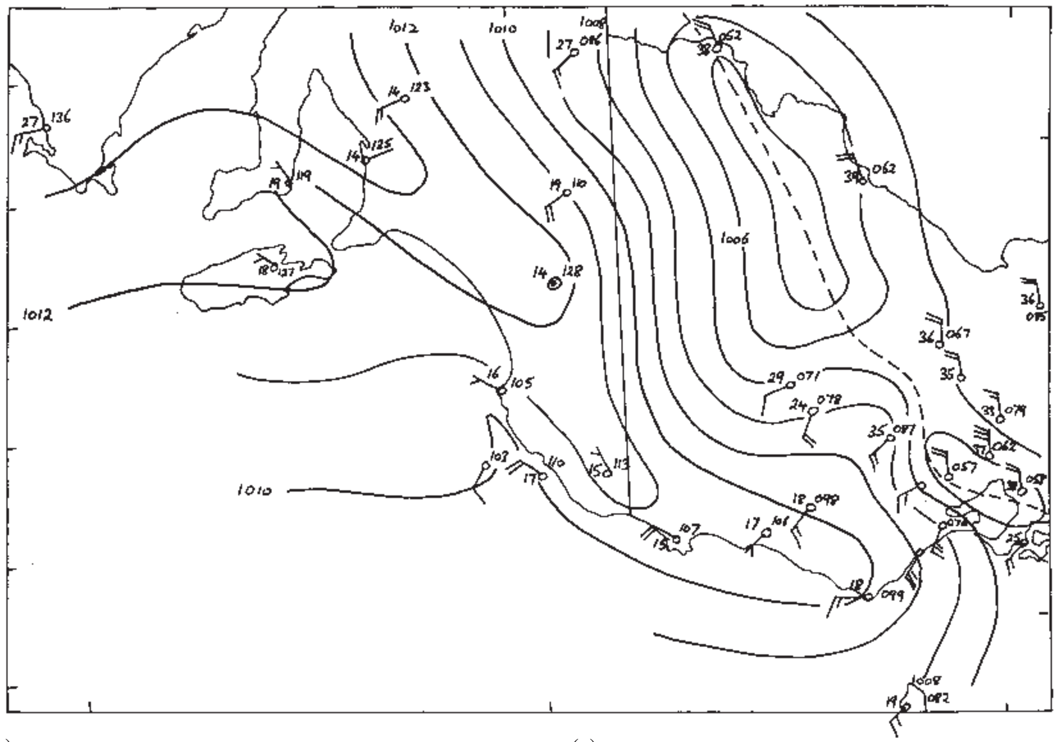


(Figs 8(a), 8(b)) show the axis of the low centred over the Otway Ranges. The inland front has progressed eastwards through Victoria, and stretches from the Victoria – New South Wales border to the coast, with marked wind shear from north-northwest to west-northwest across the thermal gradient (Fig. 8(c)). Along the coast, the model indicates the commencement of a southwesterly wind surge to the east of Cape Otway, something that was only hinted at in the forecast three hours earlier, and difficult to resolve from the observations.

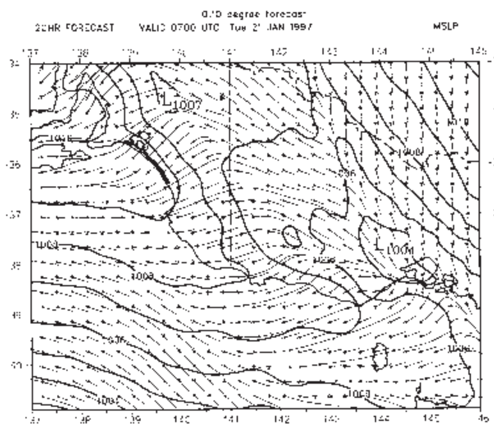
By 0700 UTC, observations (Fig. 9(a)) indicate cooler southwesterly winds through southeast SA and southwestern Victoria. The inland front has moved eastwards, to extend from near Mildura to west of Melbourne and then links with the coastal surge that

Fig. 9 As Fig. 5 for 0700 UTC 21 January 1997.

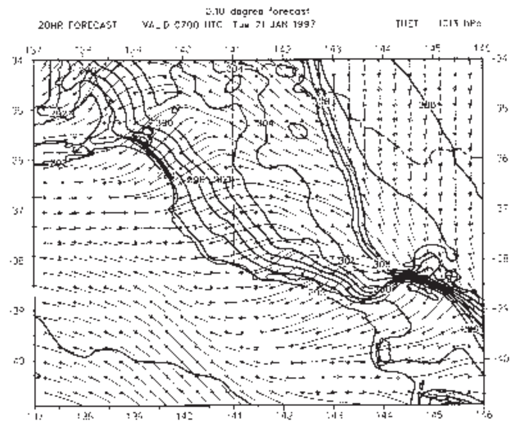
(a)



(b)



(c)



lies across Port Phillip Bay. There is a clear discontinuity in the orientation of this wind change line, also reflected in the model forecasts (Figs 9(b), 9(c)) where the effects of the surging change along the coast merge with the remarkably straight (modelled) inland front. The position of the inland front is well forecast, while the movement of the southwesterly surge along the coast is a little too slow being only at the entrance of Port Phillip Bay at this time. However the simulation of the structure and behaviour of this system is excellent.

Overall, then, this simulation of the event captures many of the observed features, particularly the developing inland front through southeastern SA, and provides an outstanding simulation of the southwesterly surge along the coast. There are some aspects of the forecast which could be better, notably the lack of sufficient backing of the winds to the west of the pre-frontal trough, and this must remain an avenue of investigation for the model developers. Using the model forecast as a conceptual model on which to base the subjective surface analysis provides an analysis which is much closer to the model forecasts than were achieved in the Victorian RFC, as these operational analyses are very much more geostrophic than are the re-analyses in this paper. This issue will be discussed further in the next two sections of the paper.

The concept of identifying the cool-change as the warm-air edge of a zone of enhance thermal gradient does well match the resolved wind change after the development of the pre-frontal trough, and after the development of the southerly surge along the coastline. Indeed, in the model, and perhaps in the real atmosphere, this is a more sensitive indicator of the 'cool change' than is the line of the wind change in the early stages of the pre-frontal trough development (see Fig. 7(c)). However, it does not well-resolve the position of the wind change over the ocean as it approaches the coast, although the faithful reproduction of the observed coastal frontogenesis suggests that the model fields contain an accurate representation of the flow. In the next section the structure of the atmosphere over the ocean, and the development of the inland front over the southeast of SA will be examined.

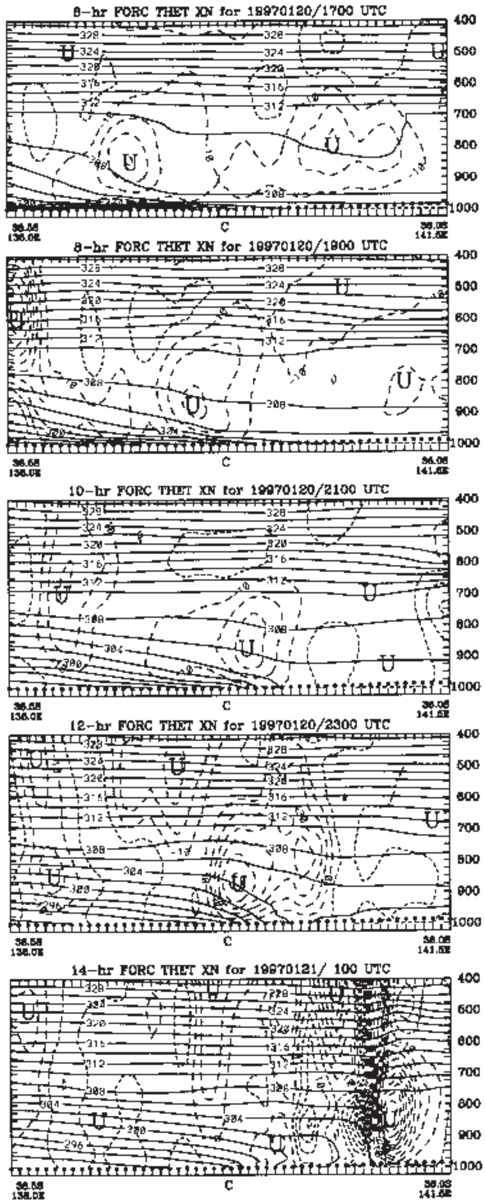
The extremely accurate prediction of the development and movement of the coastal wind surge is one of the outstanding features of this forecast. Using the model fields the processes leading to the development of this surge will be examined.

Finally, the discontinuity in the orientation of the cool-change line at the coast, and the separation in time of the development of the inland front and the coastal surge indicates that they may have different characteristics, and these may have some implications for forecasters. The vertical structures and temporal evolution of these structures will be described.

The structure of the front over the sea

It was seen in Figs 5(a), 6(a), for example, that while a wind change and a weak pressure trough could be discerned in the forecast fields while the change was offshore, little in the way of associated thermal gradient could be identified over the ocean (Figs 5(c), 6(c)). To examine this issue in more detail, a series of cross-sections of potential temperature and vertical motion normal to the southeast coast of SA are shown in Fig. 10 at two-hourly intervals starting while the front is well out over the sea. (The location of the

Fig. 10 **Cross-sections normal to the southeast coast of South Australia at 1700, 1900, 2100 and 2300 UTC 20 January, and 0100 UTC 21 January 1997. Contours are potential temperature (full lines), contour interval 2 K, and vertical motion (dashed lines, negative values long dashes), contour interval 10 hPa h^{-1}. Ascent maxima marked with U. The coastline (C) is in the centre of the section. The line in Fig. 5(c) shows the cross-section location.**



cross-section is shown in Fig. 5(c).) Looking at the upper panel, valid at 1700 UTC, a stable layer is seen over the sea extending from the surface to around 950 hPa. However, above this stable layer, in the southwest (left-hand) third of the section, is a zone of upward-sloping isentropes and an associated upward vertical motion cell – the signature of a cold front. Over the land the lowest 50 hPa is also stable, but significantly less-so than over the ocean, while right across the section, above the stable layer exists a mixed layer extending to around 700 hPa.

Back trajectories taken at points along the section show that at 900 hPa (Fig. 11) the front seen in the cross-sections is formed in the confluent flow of air from over the interior of SA (the warm side, and represented by the mixed layer air seen in Fig. 10 at 1700 UTC), and cooler air from the ocean over the Great Australian Bight. It is seen that immediately ahead of the front, air parcels from over the land have trajectories over the ocean. These parcels are cooled from below forming the stable layer seen in Fig. 10 – an internal boundary layer (Garratt 1990) formed as hot continental air is cooled as it moves southwards over the ocean. Thus, while the wind field and pressure trough at the surface are identifiable over the ocean at 10 m in Figs 5 and 6, the relation with the thermal pattern is masked by this boundary layer cooling. However, at 900 hPa (Fig. 12), there is a clear relation between the position of the surface wind change and the warm edge of the zone of enhanced thermal gradient marking the cold front. Over the ocean it is easier to identify the structure and location of the cool change, as defined by the thermal gradient, at this level than it is close to the surface.

Returning to the time series of cross-sections (Fig. 10), the modification of the front as it crosses the coast at around 2200 UTC is clearly seen. The upward vertical motion centres at 900-800 hPa can be interpreted as marking the position of the upward branch of the cross-frontal ageostrophic circulation. The close association of this ascent zone with the commencement of the cooling (upward slope of the isentropes) is readily seen. This frontal circulation moves steadily eastwards, and after crossing the coast begins to intensify. Sunrise at that time of year is close to 2100 UTC in southeastern SA, and the effect of the diurnal heating cycle over land can be seen in Fig. 10. In the first three panels, prior to sunrise, the terrestrial boundary layer cools (see the rising of the 308K isentrope), but after sunrise the surface rapidly warms, and a mixed layer to 700 hPa is established by 0100 UTC (Fig. 10). Thus the front moves from the sea to land at around the same time as the land surface commences heating. The static stability of the air ahead of the front decreases, the intensity of the cross-frontal circulation increases, the isentropes become very steeply sloped, the intensity of the vertical circulation is greatly enhanced, and the thermal gradient is enhanced – i.e. frontogenesis occurs. Finally, the hitherto extremely regular eastward progression of the frontal ascent maximum accelerates.

This simulation shows a very strong resemblance to the idealised modelling study of Reeder (1986), where it was shown that a front moving into a well-mixed daytime boundary layer showed enhanced vertical motion and some acceleration of the front after it moved into the heated onshore boundary layer.

Fig. 11 Three-dimensional back trajectory plots prior to the cool change reaching the southeast SA coast. The trajectories are calculated backwards from 990 hPa (top) and 900 hPa (bottom) for nine hours. The numbers show the pressure level at the start of the trajectory (1100 UTC 20 January 1997).

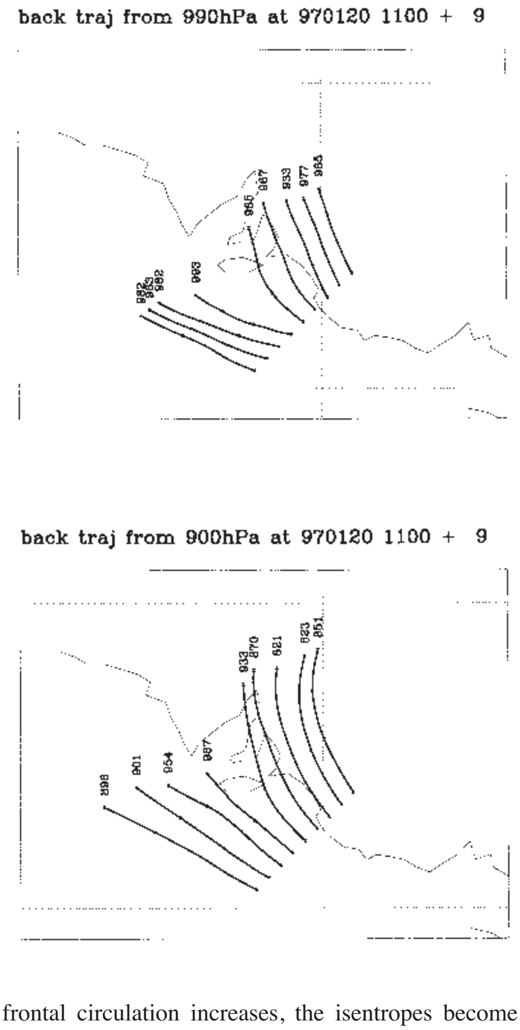
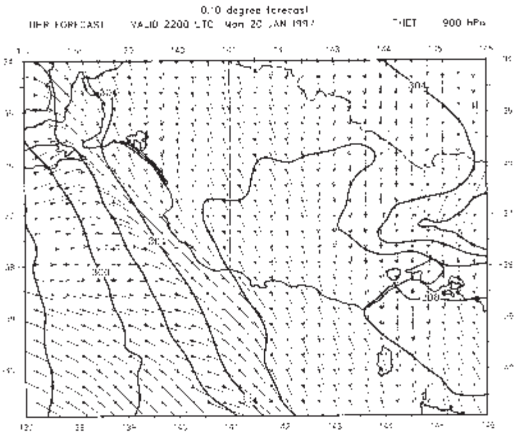


Fig. 12 The forecast 900 hPa potential temperature field (contour interval 2K) at 2200 UTC 20 January 1997 overlaid on the 10 m streamline field.



Bluestein (1986, pp. 188-192) shows from an analysis of the geostrophic momentum approximation to the Sawyer-Eliassen equation for the cross-frontal ageostrophic circulation that a weaker static stability favours stronger vertical circulations, as seen qualitatively in this case.

However, the development of the frontal zone over southeastern SA is under the control of the free-atmosphere pre-existing cool change. While this front is seen to strengthen, and to accelerate to some small extent, once it reaches land, it is not a case of 'frontal leaping', but rather a case of frontal intensification in response to boundary condition changes. It is the presence of the maritime internal boundary layer that masks the surface manifestation of the front over the ocean.

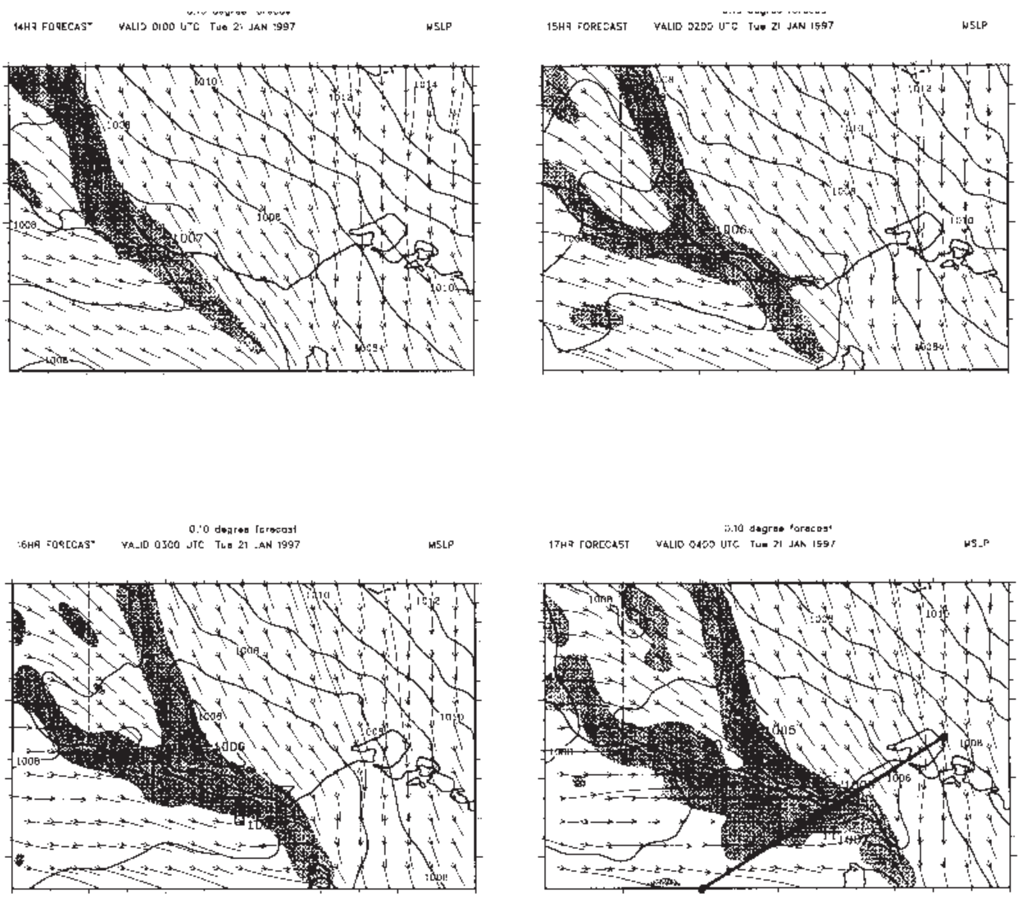
The surge along the Otway coast

It was shown in the comparison of the model forecasts and the re-analysis that a low-pressure centre formed over western Victoria during the morning of 21 January, as the trough moved eastwards. In Fig. 13, the hourly sequence of MSLP and 10 m streamlines are shown over western Victoria in the period leading up to the surge along the coast east of Cape Otway. The lowest pressure forms over land ahead of the change in wind direction, and thus leads to a reversal of the pressure gradient from offshore to onshore in the lowest levels. Thus, from a more-or-less geostrophic wind balance, with the pressure gradient directed to the southwest, and the wind northwesterly, over a period of only one or two hours the balance is disrupted. The pressure gradient force along the coast reverses, reinforcing the Coriolis acceleration (always directed at right angles to the left of the wind vector in the southern hemisphere), and a rapid backing of the wind results. With an internal boundary layer already established over the ocean as the hot offshore winds are cooled in the lowest levels, this rapid backing advects this internal boundary layer along and across the coastline, leading to rapid frontogenesis and 'surging' of the front along the coast.

Under conditions of geostrophic balance, the scalar product of the pressure gradient force and the Coriolis force vectors will be negative. However, if the pressure gradient force reverses so that it is directed to the left of the wind vector, as is the case in this event, then this product will be positive (angle between the vectors less than 90°). Figure 13 shows shaded areas where this dot product is positive. There are significant areas of positive abaric forcing (positive dot product) just to the rear of the wind change line, as would be expected. More significantly, there are also areas along the coastline of positive abaric forcing in the northwesterly flow ahead of the wind change. This is particularly apparent in a narrow band along the western coast of Cape Otway at 0200 UTC. In these regions, there is a forcing for rapid backing of the northwesterly flow, and thus for the wind change to 'surge' along the coast.

This reversal of the pressure gradient is apparently forced by surface heating, and is confined to the lowest levels of the troposphere. As the pressure trough propagates eastwards it is an initially open wave at the surface, with the pressure gradient directed (south)westwards, but as it reaches the Victorian landmass, combined with the onset of the diurnal heating cycle, and, it must be added, with a pre-existing very hot air mass over southeastern Australia, the low deepens, and, combined with the eastward advection of the cooler air in the maritime internal boundary layer, a local eastward-directed pressure gradient forms. However, this is a shallow phenomenon, and is not seen as low as the 900 hPa level, indicating that the processes leading to the surging are shallow, and partly explaining why these fronts are frequently very shallow. Indeed, the depth of the maritime internal boundary layer probably controls the depth of this surge. This is graphically indicated in Fig. 14, which shows the east-west height profile through western Victoria at 1-hourly intervals from 2300 UTC 20 January to 0500 UTC 21 January 1997. At 900 hPa, the height gradient is quite linear from east to west. However, at 1000 hPa, the trough is scarcely percep-

Fig. 13 Mesoscale model forecasts valid 0100, 0200, 0300 and 0400 UTC 21 January 1997 of 10 m streamlines, mean sea-level pressure (contour interval 1 hPa), and shaded areas where the scalar product of the Coriolis and pressure-gradient forces is greater than zero (see text for details).



tible at 2300 UTC, but deepens markedly as time progresses, particularly in the final three hours, when the pressure gradient just to the west of the trough axis becomes more marked. This is an effect of lowering pressure on the trough axis, rather than the increasing of pressure to its west. The highest height values on the western section of the section remain in the range 65-70 m throughout the six-hour period, while the height on the axis of the trough lowers from around 60 m at 0100 UTC to 44 m at 0500 UTC.

The surging of the wind along the coast east of Cape Otway is particularly interesting. A cross-section parallel to the eastern side of the Otway Peninsula (Fig. 15, see line of section in Fig. 13) shows the development of the surge. At 0100 UTC the change is well west of Cape Otway (marked C). East of the cape, the air is fully mixed down to the surface, having very little fetch to develop an internal boundary layer. West of the cape, however, there is an increasingly deep cool layer. As time progresses, the air over the ocean just east of Cape Otway continues to be that of the continental boundary layer, reaching around 308K. The eastward advection of cooler air west of the cape, enhanced by the turning of the winds due to the reversal of the coastal pressure gradient, gradually intensifies the thermal gradient off Cape Otway. Thus, when the thermal gradient reaches its maximum along the west coast of the Otway Peninsula (see Fig. 15 at 0600 UTC), the density contrast will be similar along the eastern coastline and along a parallel line just inland. However, friction over the sea will be much less than over the land. In addition, diagnostic calculations (not shown) indicate that once the front moves inland over Cape Otway

Fig. 14 **East-west height profiles at 900 and 1000 hPa along 38.5°S, from 139°E to 146°E. Units of height are m. Times shown are 2300 UTC 20 January 1997, and 0100, 0300 and 0500 UTC 21 January 1997.**

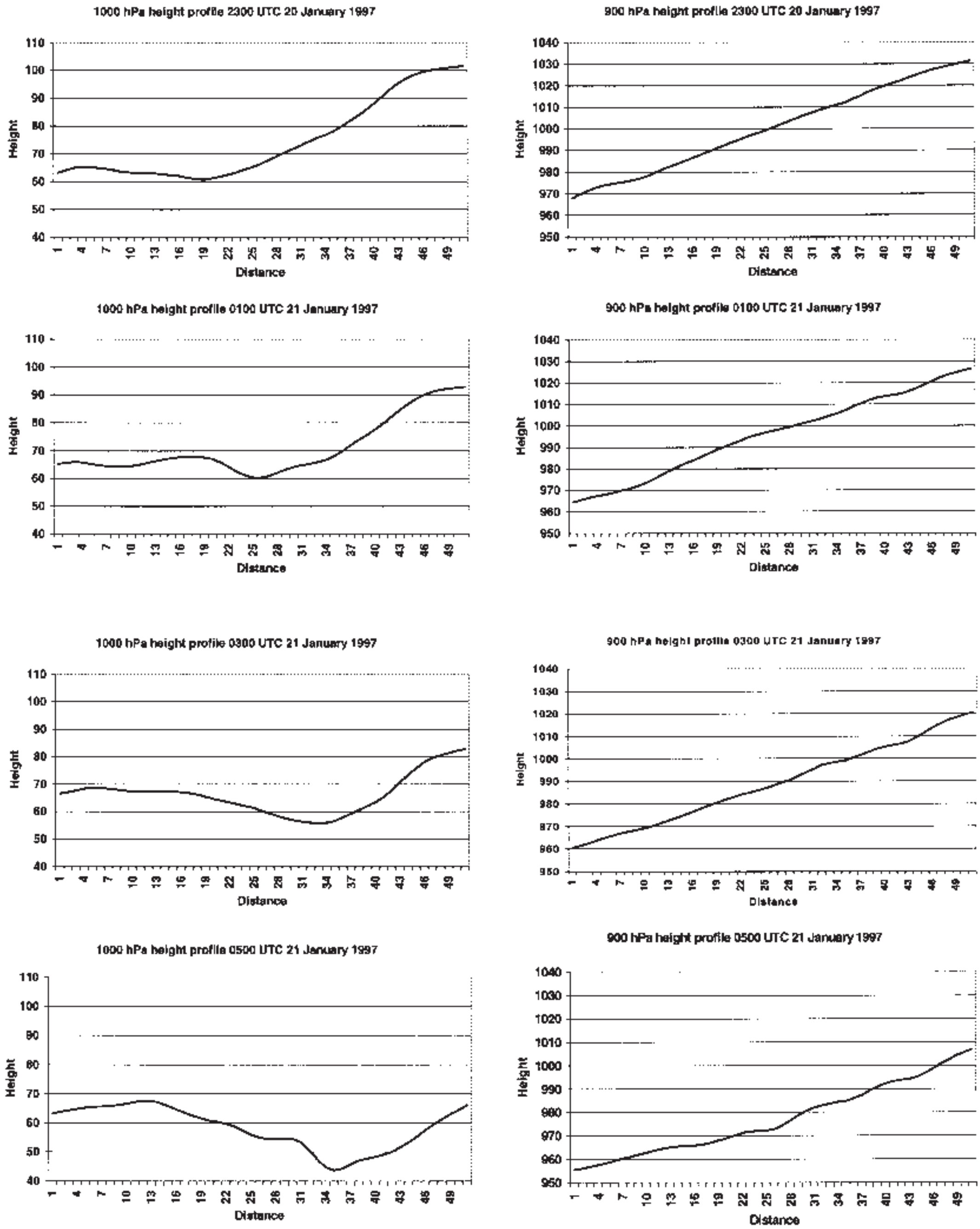
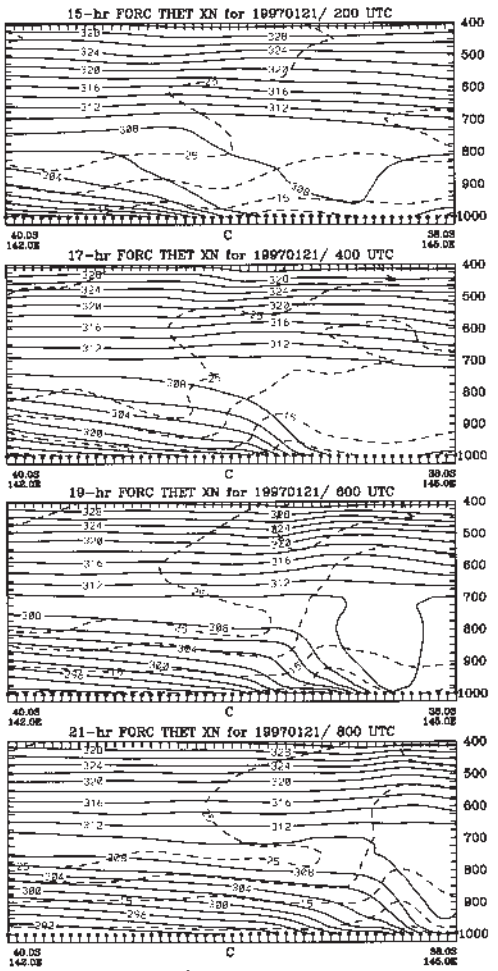


Fig. 15 Cross-sections from the mesoscale model forecast parallel to the eastern coastline of the Otway Peninsula at 0200, 0400, 0600 and 0800 UTC 21 January 1997. Contours are potential temperature, contour interval 2 K, and wind speed (dashed contours, contour interval 5 m s^{-1}). The line in Fig. 13 shows the cross-section location. The position abeam Cape Otway is marked by the C.



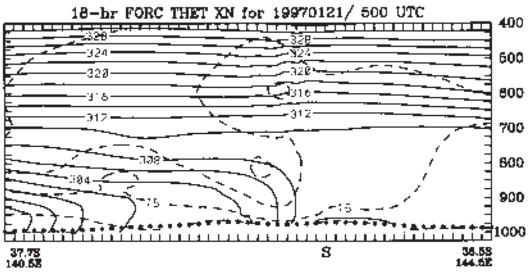
there is strong post-frontal diabatic heating, which acts to weaken the thermal gradient. There are thus multiple processes acting to enhance the movement of the change along the coastline and they appear to be very similar to those proposed by Garratt (1986).

The cross-sections (Fig. 15) show some of the characteristics of a gravity current, with the strongest winds to the rear of the 'nose'. The rate of propagation of this change along the eastern coast of the Otway Peninsula is calculated to be around 8-10 m s^{-1}. This is significantly less than the wind speed a little above the surface just to the rear of the head of the surge, in keeping with a gravity current hypothesis.

Comparative structure of the inland and coastal changes

The discontinuities in orientation and rate of movement of the change between the overland part of its movement through Victoria and its surge along the coast are also reflected in different vertical wind structures in the two areas. Fig. 15 shows the potential temperature/wind speed cross-sections along the coast, while Fig.16 shows, just for one time period as an example, the cross-section normal to the inland front through western Victoria at 0500 UTC. The coastal front shows a relative minimum near-surface wind speed just at the position of the change, with the strongest low-level winds some distance to the rear of the change. In contrast, the inland front shows a clear low-level jet at the top of the abrupt thermal gradient, around 1-1.5 km above the ground, and bears some similarities to the structure shown in Reeder and Smith's (1987) idealised modelling studies (their Figs 4,5). The level of this jet matches the height of the upper, southwest-directed, branch of the cross-frontal ageostrophic circulation that is acting to accelerate the northerly winds here. These different vertical structures of the two segments of the one 'cool change' would have considerable impact on the forecast winds before and after the change. Figure 17 shows wind direction, speed and maximum gust time series from Aireys Inlet, on the coast and representative of the coastal surge, and from Stawell, representative of the inland cool change. At Stawell the highest wind speeds occur before and at the wind change which, after the initial direction change at 0530 UTC, shows a relatively slow and continuous backing, with the wind speeds dropping after the change. At Airey's Inlet, however, while there is a sustained period of strong, gusty northerly winds before the change, wind speeds drop before and through the change before increasing for a period of around two hours some two hours after the passage of the wind change. These observed variations are in close agreement with the trends which would be expected from a space-to-time conversion from the cross-sections, and indicate that the model is not only simulating the variations in speed of movement of the change, but also resolving many of

Fig. 16 A cross-section normal to the inland change at 0500 UTC 21 January 1997. Contours show potential temperature (contour interval 2K) and wind speed (dashed contours, contour interval 5 m s^{-1}). The location of Stawell is marked with an S.



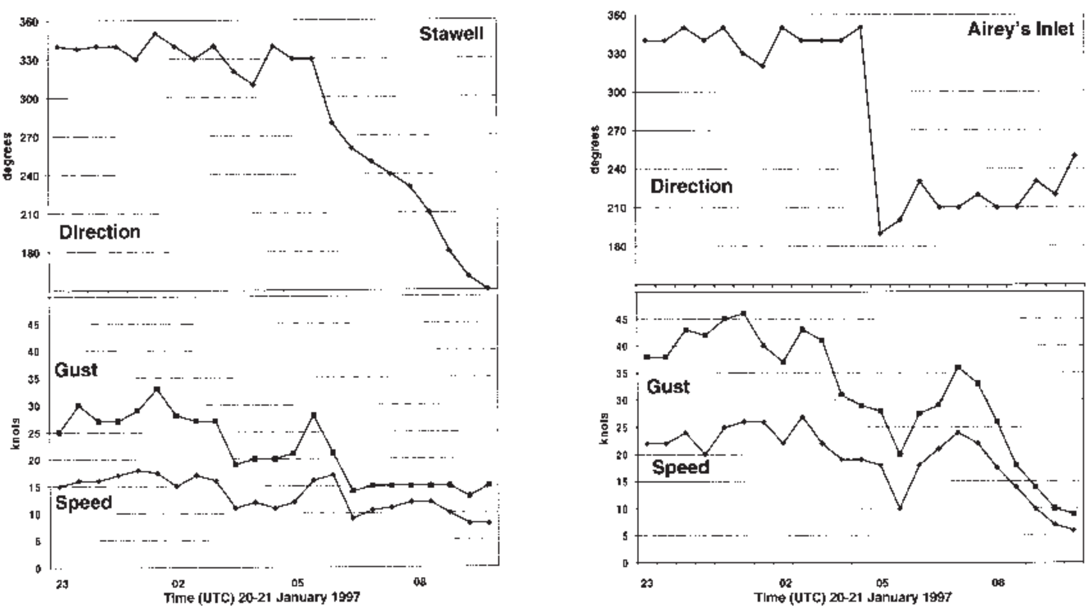
the structural variations along the change.

Discussion

This study has revealed a number of features with implications for the understanding of and the forecasting of fronts as they cross the southern Australian coastline. A pervasive factor is the role of the maritime internal boundary layer that is generated in offshore flow ahead of the approaching cold front. The fact that the surface change is difficult to find over the sea south of central SA is due to the surface modification of the thermal contrast across the front. In northwesterly flow ahead of a cold front south of SA, an internal boundary layer is established with near-surface temperatures approaching that of the sea surface. This case suggests that while a weak surface wind change and a pressure trough can be detected near the surface, the thermal contrast marking the cool change is best observed above this internal boundary layer. Further, this thermal gradient is the feature that best marks the front. Any design of frontal reconnaissance missions using manned or autonomous aircraft should bear this in mind.

The development of the internal boundary layer in offshore flow over the ocean has varying and complex effects on the movement of cool-changes along the Victorian coastline. Ahead of the change, in western Victoria, the offshore boundary layer cools ahead of the wind change. This resulting density contrast produces a pressure gradient anomaly directed onshore, and causes the northwesterly winds in this region to back before the change arrives. With this backing occurring in a pre-conditioned maritime internal boundary layer, the onshore pressure gradient anomaly is enhanced, causing the change to appear to 'surge' along the west coast of Victoria, as seen in Figs 5(b), 6(b). Once diurnal

Fig. 17 Time series of wind speed, direction and gust reports obtained from the metar observations, from 2300 UTC 20 January to 1000 UTC 21 January 1997 at (a) Stawell Airport (37.06°S 142.74°E) and (b) Airey's Inlet (38.46°S 144.09°E). Wind speeds and gusts reported in knots; wind direction in degrees.



heating commences, this process is accentuated, with pressures falling over land. This then leads to the development of a strong backing of the winds on the west coast of the Otway Peninsula – a further surge eastwards of the front, and the development of a deep differential gradient in IBL depth at Cape Otway around 0200 UTC. The abrupt change of coastal orientation at the cape then allows the rapid development of a gravity current-like surge along the east coast of the Otway Peninsula that moves well ahead of the longitudinal position expected from an extrapolation of the orientation of the inland front.

In the early stages of development of the inland front the warm-air edge of the thermal gradient provides a better early indication of the position of the developing cool change in the model output than does the wind field. The direction shear at this cool change only becomes obvious as the wind backs with time to the west of this line. While it is acknowledged that the model winds back too slowly with time, this process may be part of the reason why fronts were sometimes considered to 'leap' across the coast. It may be that the directional wind shear at the front was not particularly obvious until after the thermal gradient intensified and time had allowed sufficient backing to occur. This points to the benefits of using both the thermal and the wind fields from mesoscale model output when identifying change lines in mesoscale NWP output.

While the main frontal change is seen to be generally under the 'control' of the 900 hPa change, which propagates eastwards relatively continuously, the surface manifestation of this change is strongly modified by local land-sea heating and planetary boundary layer (PBL) structure contrasts. This leads to rapid evolution in time and space of frontal intensity, movement and structure. Of particular note are the differing vertical wind structures relative to the cool change line seen in the inland front and the coastal surge parts of the change. In the inland case, the strongest winds at the surface are both predicted, and were observed, to be ahead of the change, with a relatively rapid drop in speed with time after the change. However, along the coast there was an additional surge of strong winds after the change. The understanding and predicting of these differences are crucial for forecasts for fire suppression operations. These differing 'types' of changes are well illustrated by comparing the anemometer traces for the Ash Wednesday 1983 and the Black Friday 1939 cool changes (see Bureau of Meteorology, 1984, Figs 77 and 91). In each case there were strong, gusty northerly winds ahead of the change, but in 1939 the wind speed dropped rapidly after the change, while in 1983 there was a sustained period of strong winds following the change.

The process of attempting to fit the verifying MSLP analyses to the patterns forecast by the mesoscale model is quite instructive. In a nowcasting mode, the observations of wind and temperature along the Victorian coast are interpreted differently if the trough is analysed as an open wave, as is the case if a moderate degree of geostrophy is assumed in the analysis. Such an analysis, though, would not imply any dynamic basis for deducing an imminent coastal surge. If the analysis is based on the numerical model (see Figs 5-9), then highly ageostrophic analyses result, and the observations at, say, 0100 UTC then support the mesoscale model forecast, and a higher degree of confidence could be placed in the later parts of the NWP forecast. It is clear that the observational network is insufficiently dense to unambiguously define patterns and circulations such as those depicted in this mesoscale NWP forecast, and so this process could be seen as a paradigm for mesoscale analysis in an RFC. Numerical weather prediction system forecasts are not always as accurate as that seen in this paper, but the subjective mesoscale analyses could be adjusted in phase and amplitude to use the dynamic process information inherent in the model forecasts.

Many features of previously published conceptual models of fronts over the southeast of Australia have been identified in this case study, and are shown to be valuable descriptions of the processes seen in this case study. However, it must also be remembered that the resulting processes and evolution are strongly dependent on the flow configuration and its relation to coastline orientation, and on the phase of the larger-scale trough relative to the diurnal heating cycle. Thus each front will behave differently. It has been shown, though, that many different processes may operate in different parts of the front, or in different phases of its evolution. If nothing else, this study shows that a single conceptual model does not fit even one front for more than a short length of time, and indeed, over relatively short periods. Mesoscale NWP model output provides the framework to allow for this variation, and to incorporate this information into the forecast process.

Acknowledgments

The author would like to thank Tony Bannister, of the Victorian Regional Office, for the provision of archived metar data used in the verifying analyses, and also in Fig. 17. As always, such a study as this can only be achieved with a state of the art NWP model, and the efforts of Kamal Puri, Noel Davidson, Robin Bowen and Les Logan in the development of the

LAPS system are gratefully acknowledged.

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