A Numerical Study of a Southeast Australian Coastal Ridging Event

K. J. Tory*

Environmental Studies, University of Northern British Columbia, Prince George, British Columbia, Canada

C. J. C. Reason

School of Earth Sciences, University of Melbourne, Parkville, Victoria, Australia

P. L. Jackson

Environmental Studies, University of Northern British Columbia, Prince George, British Columbia, Canada

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ABSTRACT

A numerical study of the 9–11 November 1982 southeast Australian coastal ridging event is presented. The mesoscale coastal features of this event have been previously described as a coastally trapped disturbance (CTD). However, the analysis presented in this paper shows that the model coastal trapping is of secondary importance in generating the coastal ridging. Given the potential controversy, particular emphasis in this paper is placed on identifying the causes of the ridging. The paper follows a previous study by the authors in which a Colorado State University Regional Atmospheric Modeling System simulation of this event was validated.

In this event, contributions to ridging along the southeastern Australian coast came from both synoptic and mesoscale phenomena in the model. A Southern Ocean anticyclone that tracked east from the Great Australian Bight toward the southern Tasman Sea led to a large-scale steady pressure increase over the entire southeast Australian region. Ridging at the Victorian coastal stations developed with the passage of a synoptic-scale wave of cooler denser air at midlevels, and the arrival of low-level cooler marine air with a wind shift from the southwest to a more southerly direction. On the east coast, the pressure change was most abrupt in the south as warm continental air was replaced with cooler marine air after strong winds passed through Bass Strait and turned left around the southeast continental corner. The flow closest to the coast continued turning left, enhanced by the daytime lowering of pressure over the continent, and pushed inland to the mountain barrier.

The most intense part of the wind surge crossed the coast during the evening near the Hunter Valley (a significant gap in the east coast mountain barrier). This led to flow splitting inland up the Hunter Valley and northward along the east coast. The coastal component developed into a CTD in the form of a wind surge that accelerated ahead of the region of onshore forcing until it encountered a reduced mountain barrier farther north, and the flow spilled inland signaling the end of the model event.

1. Introduction

Coastal ridging is frequently observed in the southeast Australian region during the warmer months and is linked to the passage of midlatitude Rossby waves. More often than not the ridging is preceded by the passage of a cold front (Speer and Leslie 1997) that forms in the deformation region between the anticyclone and a trough to the southeast. During the austral summer, the subtropical ridge is typically centered just south of the Australian continent and each anticyclone propagates from west to east with a period of about 5–7 days. As the anticyclone propagates toward the east coast it first directs cooler southerly flow onto the south coast of Victoria (see Fig. 1 for locations) and then up the east coast where the flow may be blocked by or trapped against the Great Dividing Range (GDR) provided sufficient vertical stability exists. Here, we choose to make the following distinction between blocked and trapped flow. Blocked flow is flow dammed up behind the mountain barrier, and trapped flow is blocked flow that turns and begins propagating along the coast in a manner satisfying the definition of a coastally trapped distur-

* Current affiliation: Bureau of Meteorology Research Centre, Melbourne, Victoria, Australia.

Corresponding author address: Kevin Tory, Bureau of Meteorology Research Centre, GPO Box 1289K, Melbourne, Victoria 3001, Australia.
E-mail: K.Tory@bom.gov.au

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It is possible that the depth of the front and the track of the subsequent anticyclone are such that the GDR does not present an effective barrier to the postfrontal cold air and hence the coastal ridging is not trapped. While a not insignificant fraction of category 1 ridging events involve some type of mesoscale trapped component (M. S. Speer 1998, personal communication), the contribution of CTD events to southeast Australian coastal ridging remains to be accurately determined.

The typical North American CTD is generated following offshore flow along the northern Californian coast and, to the south, onshore flow that is blocked by the coastal mountain barrier (Dorman 1985, 1987; Mass and Albright 1987; Thompson et al. 1997; Ralph et al. 1998, 2000; Skamarock et al. 1999; Guan et al. 1998; Jackson et al. 1999). Convergence of the onshore flow on the barrier leads to localized raising of the marine
boundary layer (MBL) and an associated alongshore pressure gradient generates an ageostrophic flow along the coast, which is trapped laterally by the mountain barrier and vertically by stable stratification.

We suspect there would be greater difficulty in separating blocking from trapping in the Australian category 1 ridging events, compared with the North American events, due to the relative propagation speeds of the forcing mechanisms that lead to blocking. The blocking that initiates the typical North American CTD is generated by the relatively slow offshore migration of a continental low, which produces onshore flow that becomes blocked by the coastal mountain barrier south of the low center. If, as proposed above, many of the southeastern Australian ridging events are caused by postfrontal flow blocked against the coastal mountain barrier, then it is reasonable to assume that the region of onshore forcing would propagate with the front, at an alongcoast velocity considerably faster than the typical North American forcing mechanism, and in the direction of any possible trapped disturbance. Intuitively it seems less likely that trapping will occur in such a situation since any trapped disturbance would need to propagate at least as fast or faster than the forcing mechanism, that is, the front. This, combined with the lower southeastern Australian coastal mountain barrier and the lack of a semipermanent MBL inversion, suggest trapping in the southeastern Australian region is less likely to occur than off the west coast of North America.

Another phenomenon that may or may not include a trapped component is the well-documented southerly buster (or burster—hereafter SB). The SB is a wind surge, with gusts greater than 15 m s$^{-1}$, generated by the interaction of a cold front with the mountain barrier. They are typically accompanied by sudden temperature changes and low cloud, but rarely significant rain (Colquhoun 1981). A detailed description of the SB is given by Reid and Leslie (1999). Although sometimes loosely classed as a type of CTD, McInnes and McBride (1993) note that the SB does not always fit the definition of the trapped disturbance proposed above. They mention that although the Baines (1980) theoretical model of the SB predicts a trapped component leading the synoptic front (a density current flowing along the topography with exponential decay away from the topography), the observations taken during the Southerly Buster Observational Program experiment show no sign of exponential decay. However, their modeling results do show signs of a trapped gravity current on the south coast but no signs on the east coast (McBride and McInnes 1993). Colquhoun (1981) studied coastal wind and pressure traces at a number of east coast stations during strong SB events (greater than 20 m s$^{-1}$) between January 1972 and January 1978. He concluded that the strong southerly changes generally developed ahead of the fronts. In the absence of any across-shore observations it is not possible to conclude whether or not these leading southerly surges were trapped, although their position ahead of the main front is consistent with trapping in the Baines (1980) model.

Although numerous observational studies of North American CTDs (Dorman 1985, 1987, 1988; Mass and Albright 1987, 1988; Reason and Dunkley 1993; Ralph et al. 1998, 2000) and numerical simulations of two events (Thompson et al. 1997; Guan et al. 1998; Jackson et al. 1999) have been performed, there has been relatively little research into Australian events. As a result, many aspects of Australian CTDs are not well understood. Previous research into Australian CTDs has essentially concentrated on the event of 9–11 November 1982, which propagated from near Melbourne on the south coast to about Brisbane on the east coast (Holland and Leslie 1986; Reason and Steyn 1992; Reason et al. 1999—hereafter HL86, RS92, and RTJ99, respectively). In addition to this work, Speer and Leslie (1997) have presented a climatology of the various types of synoptic and subsynoptic ridging on the east coast of Australia of which the CTD phenomenon is a subset. In their analyses of the 10–11 November 1982 coastal ridging event, HL86 and RS92 essentially had access only to coastal observations. The only offshore data available were cloud structure from satellite images, and mean sea level pressure from the somewhat subjective hand analyses. The coastal pressure, wind, and temperature data were consistent with a shallow gravity current of approximately the depth of the mountain barrier to the west, which propagated up the east coast. The handrawn mesoscale charts identified ridging in the coastal region, and the satellite images showed the development of stratus cloud that matched the apparent northward propagation of the gravity current. However, without detailed across-shore observations there was no way to confirm their CTD interpretation. HL86 proposed that onshore flow onto the Victorian coastal ranges during the November 1982 event generated a Kelvin wave–type disturbance that propagated through and ahead of a trapped gravity current. With the aid of a numerical simulation of the event using the Colorado State University Regional Atmospheric Modeling System (RAMS; Pielke et al. 1992), RTJ99 pointed out that there was insufficient vertical trapping for a Kelvin wave to form on the western Victorian coast, and suggested instead that convergence of cool southerly air against the mountain barrier in eastern Victoria provided a reservoir for a trapped gravity current–type ridging that propagated up the east coast as far as about Brisbane.

Although the 10–11 November 1982 event and the SB both fit into the Speer and Leslie (1997) category 1 ridging event, they are very different phenomena. In most of the examples presented in McInnes and McBride (1993), McInnes (1993), Reid and Leslie (1999), and Reid (2000), the front that spawned the SB was strong and formed along the dilatation axis of a synoptic-scale deformation region. They tended to penetrate far inland as well. The front, or fronts, associated with the November 1982 event were considerably weaker.
They formed hundreds of kilometers behind the dilatation axis of the synoptic-scale deformation region (see Fig. 2 of RTJ99), did not penetrate inland, and appeared to move out into the Tasman Sea well south of the location at which the coastal ridging terminated (see Fig. 4 of RTJ99). As mentioned above, both HL86 and RTJ99 proposed various forms of CTDs to explain the ridging. In this paper we propose an alternative explanation based on the RAMS simulation and a number of sensitivity simulations in which trapping is found to be of secondary importance.

The RAMS simulation of this event (RTJ99) was validated extensively against all available observational data with favorable results. RTJ99 showed that the model reproduced the coastal environment quite well, which gives confidence in the model performance away from the coast. The current study extends that of RTJ99 by examining the RAMS simulation in much greater detail in order to determine the exact cause of the coastal ridging and the associated abrupt temperature and pressure changes, and to identify any evidence of trapping. In addition, sensitivity simulations experimented with horizontal resolution and topography. They included one high-resolution simulation of 12.5 km, and simulations with the topography removed and multiplied everywhere by 1.5. Two further simulations included the mountain barrier with the Hunter Valley filled in, one with the standard topography and the other with the topography multiplied everywhere by 1.5.

In section 2 of the paper the mechanisms behind the coastal ridging are explained and comparisons with the observations are made where possible. Section 3 discusses the dynamics of the flow that led to the ridging along the east coast to the Hunter Valley, and in section 4 the model ridging, blocking, and trapping are discussed further and summarized.

2. Coastal ridging

A detailed account of the synoptic and mesoscale background has been presented in RTJ99 and will only

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**Fig. 2. Model representation of geopotential and winds at 850 hPa on grid 1 at (a) 0000 UTC 9 Nov, (b) 0000 UTC 10 Nov, (c) 1200 UTC 10 Nov, and (d) 1800 UTC 10 Nov 1982. Contour interval is 20 m. The full and half barbs represent 10 and 5 m s⁻¹, respectively.**
be briefly mentioned here. An eastward propagating anticyclone centered in the Great Australian Bight in conjunction with a trough to the south of Victoria directed westerly flow over the Victorian coastal region (Fig. 2a). As the anticyclone/trough system propagated eastward, the flow became more southerly on the Victorian coast (Fig. 2b) before southerly flow began to extend up the New South Wales (Fig. 2c) and southern Queensland coasts (Fig. 2d; see Fig. 1 for locations). On the east coast, the cooler flow originating from the south was advected northward in a broad band over the Tasman Sea and turned inland toward the mountain barrier. The density change associated with the replacement of warm continental air with the marine air was largely responsible for the coastal ridging. On the south coast the explanation for the ridging includes a significant contribution from midlevel waves as well. At all coastal stations the pressure change was considerably more abrupt than would be expected from the changing flow associated with the slowly evolving synoptic system described above. This pressure rise is best illustrated in Fig. 3, which shows both the model-generated (thick) and observed sea level pressure at some of the coastal stations featured in Fig. 1.

a. Victorian coastal pressure change

In this section, the contribution of synoptic advection of colder air and of midlevel waves to the surface pressure rise along the Victorian coast is examined. It will be shown that at the Victorian stations, the sharp surface pressure rise is related to the midlevel waves and associated synoptic cold advection, and the low-level displacement of the heated planetary boundary layer (PBL) with cooler marine air. In eastern coastal Victoria the convergence of this flow against the higher coastal mountain barrier contributes to the pressure rise as well.

As the winds on the south coast shifted from westerly to southerly the pressure increased by 10–15 hPa in less than 24 h (Figs. 3a–d). The model and observed traces show a very similar pressure rise pattern although the former is delayed by a few hours. A lag of a few hours in the timing of the observed pressure rise from west to east was present in the observations, which was one of the factors that led HL86 to conclude that the disturbance propagated at a rate close to 40 m s\(^{-1}\) on the south coast and was of Kelvin wave form. Figure 3 shows that the model lag was smaller than the observed lag and that, between Laverton and Gabo Island, the lag was difficult to distinguish, which, if the Kelvin wave explanation was correct, would imply propagation speeds far in excess of 40 m s\(^{-1}\). This fact together with the relatively low topography between Mount Gambier and Laverton (at most 400 m tall) and the depth of the stable layer near 1500 m as shown in both the model (Fig. 4a) and observed (Fig. 6a of RTJ99) Laverton time–height potential temperature profile leads one to question the HL86 interpretation for that part of the south coast between Mount Gambier and just west of Sale. Clearly, favorable conditions for trapping and blocking were not present there and another explanation for the model pressure rise is necessary in this region.

An alternative explanation of the surface pressure rise at the south coast stations is best illustrated in Fig. 4a, which shows a time–height contour plot of the model potential temperature at Laverton (La; Fig. 1) with wind barbs overlayed and the sea level pressure time series below. It can be seen that the pressure rise commenced soon after 0000 UTC 10 November, which roughly coincides with the wind shift to a more southerly direction between 1000 and 400 hPa. Furthermore, the isentropes between 825 and 575 hPa lifted substantially between 2000 and 0400 UTC 9 and 10 November, respectively, before subsiding again. This midlevel wave is also evident at all other south coast stations (e.g., Fig. 4b). The time lag between the commencement of the rising isentropes and the rise in pressure is due to diurnal effects opposing the pressure rise between 2000 and 0200 UTC (0600 to midday LST). This is illustrated in Fig. 4c, which is similar to Fig. 4a except change in density is contoured. The time evolution of the vertical profile of the density change is useful for determining which part of the atmosphere is responsible for changes in surface pressure. For this reason the entire model domain has been included, so that surface pressure changes associated with changes in the atmosphere above 400 hPa can be accounted for.

Figure 4c shows a diurnal oscillation of increasing and decreasing density below about 880 hPa associated with daytime heating and nocturnal cooling of the PBL. The advection of cooler and denser air accompanying the low-level wind shift near 0400 UTC 10 November brought about a density increase of similar magnitude and depth as the nocturnal cooling provided on the previous night. Figure 4a shows that the night following the change was only about 2°C cooler than the night prior to the change, which suggests that the low-level temperature change was relatively weak and associated pressure increases would not have been much greater than the diurnal changes of the previous day. At the midlevels (about 560–880 hPa) the density increases between 2000 UTC 9 November and 0600 UTC 10 November and decreases again afterwards. This is coincident with the midlevel wave mentioned above. An examination of the observed fields equivalent to Fig. 4a at Laverton (Fig. 4a of RTJ99) shows consistency between the model results and observations with the main difference being the larger amplitude and earlier timing of the observed midlevel wave. The timing of the low-level wind shift is consistent between the model and observed fields, which, given the inconsistency of the timing at midlevels, suggests the low-level wind shift develops in response to diurnal influences and is not necessarily associated with the midlevel waves. (In both the model and observation, the change occurs in the early afternoon.)
Using a hydrostatic argument, the initial pressure increase at the south coast stations can be accounted for by the addition of a relatively dense air mass through either cold advection of marine air accompanying the southerly wind shift and/or lifting through vertical advection. A pressure increase due to the second mechanism may be related to the midlevel wave in the isentropes evident in both the observations (Fig. 6a of RTJ99) and the model (Fig. 4a). An examination of the grid 1 potential temperature and winds at 3288 m (not shown) indicates a northward advection over the region of cooler air originating from relatively high latitudes to the south of Australia. This advection represents the flow around the trailing edge of the midlevel trough and associated frontal system seen in the synoptic chart (Fig. 2a of RTJ99). As the southerly component of this ad-
vection increases late on 9 November and early on 10 November with eastward passage of the trough, the midlevel isentropes rise. A few hours later relatively warmer air is advec ted over the region so that the isentropes subside. At the surface however, pressure continues to rise as the anticyclone in the bight tracks eastward (see the increasing density in Fig. 4c throughout most of 10 November between about 300 and 200 hPa associated with the rising tropopause) and the density continues to increase at low levels.

In summary, there are three regions of increasing density evident in Fig. 4c during the period of increasing surface pressure. At low levels the increasing density was associated with the advection of a slightly cooler air mass accompanying the wind shift and the nocturnal part of the diurnal cycle. At midlevels the increasing density was associated with the arrival of the midlevel waves, and at upper levels the more gradual increase was associated with the arrival of the high pressure system. The magnitude and temporal extent of the density changes evident in Fig. 4c suggest that the low- and midlevel mechanisms contribute equally to the surface pressure change.

Figure 4b, a time–height contour plot at Orbost in eastern coastal Victoria (about 300 km from Laverton), contains the same midlevel wave in the model isentropes although the maximum amplitude occurs about 2 h later near 0600 UTC. This lag is linked to the eastward propagation of the trough. Unlike Laverton there is a clear temperature change exceeding the diurnal variation (minimum temperatures were 4° and 6°C cooler than the previous two nights), which was associated with the wind shift to a more southerly direction. By comparison with Laverton, the model topography is significantly
higher (1000–1300 m) north of Orbost and blocking of the low-level southwesterly flow of cold air on 10 November (Fig. 4b) is more likely than at Laverton.

It appears that at central and western Victorian stations (e.g., Laverton) the sharp surface pressure rise is related to the midlevel waves and associated synoptic advection of colder air (see Fig. 4c between 850 and 600 hPa, 1800–0400 UTC 9–10 Nov), and the low-level wind shift and the displacement of the heated (low density) PBL with cooler marine air. In eastern coastal Victoria (e.g., Orbost, Fig. 4b), convergence of the low-level southerly to southwesterly flow against the higher coastal mountain barrier is likely to make a contribution to the surface pressure rise as well. Although it is possible that a trapped component could have existed in eastern Victoria, the model results do not clearly show any trapped disturbance there.

b. New South Wales and southern Queensland coastal ridging

In this section, the importance of the turning of the region of strong winds into the south Tasman Sea and the subsequent ridging along the east coast are discussed. The results indicate that in the model there is only significant evidence for a trapped component north of the Hunter Valley.

The dominant mechanism responsible for the east coast pressure change was the onshore advection of dense marine air, which had been advected through Bass Strait and northward over the Tasman Sea before turning inland to the coastal barrier. Strong winds in the Southern Ocean southwest of Victoria and Tasmania (abbreviated to Tas. in Fig. 1, grid 1) moved through and over southeast Australia by 0000 UTC 10 November (Fig. 5a). Deep daytime convective mixing over Victoria and Tasmania retarded the flow, which resulted in a deep (about 2000 m) wind speed maximum through Bass Strait (Fig. 6). By 0400 UTC the wind maximum had propagated beyond Gabo Island (at the continental corner) and had begun to turn left in the Tasman Sea (Fig. 5b). Four hours later two regions of strong gradients in wind speed appeared, one in a broad east–west band near Nowra (indicated by an N in Fig. 5c) and the other narrower band farther south, close to the coast and at the leading edge of the shaded wind maximum in Fig. 5c. The former is the leading edge of the Bass Strait wind maximum, and the latter marks the edge of a semistationary maximum associated with flow around the topography.

The position of this semistationary maximum is largely determined by the slowly changing flow orientation through Bass Strait, and propagates around the continental corner at about 2 m s\(^{-1}\). The winds fan out and decelerate just beyond the continental corner similar to the exit region of a jet. The fanning out of the flow is strengthened near the coast by the daytime acceleration toward the continent due to surface heating lowering the pressure there. The advection of cooler air by the southerly wind strengthened the land–sea temperature contrast, which enhanced the sea breeze in both depth and intensity at the southern east coast stations and led to an earlier inland penetration to the mountain barrier. Meanwhile the strongest component of the leading wind maximum had continued to propagate up the coast while turning gradually to the left so that by 1200 UTC the direction of flow was southeasterly (Fig. 5d). During this time onshore flow, enhanced by the local sea-breeze circulation (particularly at lower levels) had propagated along the coast lagging slightly the wind maximum out in the Tasman Sea (not shown). At 1200 UTC the wind maximum was about to cross the coast near the Hunter Valley.

After crossing the coast about 4 h later, the flow had split into two streams, one through the Hunter Valley and the other to the east of the mountain barrier north of the Hunter Valley (see the two shaded regions in Fig. 5e). By 2000 UTC both streams had intensified (Fig. 5f). The first had now begun to descend down the far side of the Hunter Valley, and the second had become blocked by the mountain barrier north of the Hunter Valley. On the coastal side, the winds rotated to a more alongshore direction and a surge developed that pushed ahead of the main wind maximum, which was located farther out in the Tasman Sea (Fig. 5f).

This surge had the following qualities consistent with CTD: flow rotation to a more along-barrier direction, flow acceleration and advancement beyond the forcing region, and a surge width of the order of 100 km (i.e., approximately a Rossby radius). The across-shore vertical potential temperature structure (Fig. 7, along the section marked C in Fig. 1) shows a raised MBL against the barrier, which decays with distance from the barrier. This in itself is not sufficient evidence for trapping since blocked flow would show the same raised MBL. However, when combined with the cooler air mass against the barrier (which originated from farther south than the adjacent flow) and the evidence of the enhanced winds within 100 km of the coast (shaded), the argument is strengthened considerably. The disturbance is not as clearly defined as the typical North American CTD for two reasons: (i) it is relatively young (only a few hours

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Fig. 5. Grid 2 wind speed (contour interval 1 m s\(^{-1}\), vectors indicate wind direction) at 647 m for (a) 0000, (b) 0400, (c) 0800, (d) 1200, (e) 1600, and (f) 2000 UTC 10 Nov. The shaded areas highlight wind maxima referred to in the text. In (b) and (c) wind speeds greater than 19 m s\(^{-1}\) are shaded, and in (d)–(f) wind speeds greater than 15 m s\(^{-1}\) are shaded. The labels N and CH in (c) and (f) point to the position of the coastal towns Nowra and Coffs Harbour, respectively.
since the initiation), and (ii) unlike typical North American CTD the region of forcing is propagating at a speed similar to the trapped feature.

However, after a few hours, the disturbance had propagated beyond the tall part of the mountain barrier, flow began to spill inland, and the disturbance dissipated. This is illustrated in Fig. 8, which shows the horizontal wind speed along a vertical section marked D in Fig. 1. At 1600 UTC the wind surge core was approaching the taller mountain barrier north of the Hunter Valley (Fig. 8a). Leading the core (between 600 and 800 km) there was already evidence of flow acceleration and rotation toward an alongshore direction. Four hours later (Fig. 8b) the wind maximum had clearly intensified from 15 to 20 m s\(^{-1}\) and the stronger winds were largely confined to a depth similar to the barrier height. Figure 8b shows the trapped flow spilling over the barrier as early as 2000 UTC 10 November just north of Coffs Harbour. Four hours later there was strong inland flow over the low hills west of Brisbane and the wind surge weakened (Fig. 8c). Cold air advection accompanied the wind surge and led to a significant drop in temperature at all the coastal stations. But since the alongshore temperature gradient was relatively weak here after 1600 UTC 10 November (Fig. 9d), the temperature and pressure changes were less abrupt than at stations south of the Hunter Valley (cf. Figs. 3d–g with Figs. 3h,i).

The sharp pressure rise at the east coast stations can be attributed to the change in coastal air mass from the daytime continentally heated atmosphere to the cooler marine air mass that recently passed through Bass Strait (Fig. 9). At 0400 UTC (1400 LST) an alongshore temperature gradient existed along the entire coastline due to the land–sea heating contrast (Fig. 9a). Between Gabo Island and Nowra this gradient is enhanced immediately offshore by the cooler flow that had turned to the left and begun to flow northward over the Tasman Sea. A weaker north–south gradient exists in the Tasman Sea between about Gabo Island and Sydney, with the strongest gradient offshore of about Nowra, which roughly coincided with the northern edge of the wind surge. Four hours later (0800 UTC, 1800 LST) the alongshore temperature gradient had begun to push inland along most of the east coast due to the sea breeze (Fig. 9b). How-
ever, south of Nowra the change in temperature was enhanced by the onshore advection of the cooler and deeper air mass advected into the Tasman Sea by the Bass Strait wind maximum. At these locations, the timing of the cooler air mass arrival (within a couple of hours of the sea breeze) led to a more sudden pressure change than at stations farther north where the sea breeze arrived before the Bass Strait air mass (note the small model pressure increase at about 0600–0800 UTC due to the sea breeze, followed by the larger increase near 1400–1600 UTC at Coffs Harbour and Brisbane in Figs. 3h,i). Note also the excellent agreement of the pressure change timing and intensity between the model and observations at all the east coast stations (Figs. 3d–i), including the two-step pressure change at Brisbane (Fig. 3i). The observed two-step pressure change could not be determined at Coffs Harbour due to missing data between 1000 and 1600 UTC.

Although the onshore flow south of the Hunter Valley was strong enough to force the cooler fluid layer to ascend part way up the mountain barrier after being blocked (not shown), there was no evidence of trapping.
along that part of the coast. A third grid of 12.5-km horizontal resolution was experimented with to see if the lack of any trapped flow was due to poor resolution, but it also failed to show evidence of trapping. The timing of the onshore flow south of the Hunter Valley (early evening) may not have been particularly conducive to trapping since there may have been insufficient time for the necessary stable layer to develop on the coastal side of the mountain barrier after the deep convective mixing had subsided.

However, when the considerably stronger wind maximum from farther offshore had turned enough to make landfall near the Hunter Valley there was sufficient energy for a trapped wind surge to develop, despite splitting of some of this energy through the Hunter Valley. It is interesting to note that the observed winds at Williamtown (at the mouth of the Hunter River) show a marked peak at 1000 UTC of 14 m s\(^{-1}\) in a southeasterly direction, and southerly winds of 6 m s\(^{-1}\) 6 h later (next available observation). The true peak could have been later than 1000 UTC. This southeasterly wind maximum is remarkably consistent with the arrival of the more intense part of the model wind maximum. Furthermore, the southerly wind surge at Coffs Harbour that reached a maximum at 2000 UTC (Fig. 5f) was consistent with the observed wind maximum at the same time, although the model surge was stronger and longer lasting.

3. Bass Strait wind maximum

The previous section identified the wind maximum that passed through Bass Strait and turned left around the continental corner as the key phenomenon ultimately responsible for the coastal ridging on the east coast of the continent. In this section we show that it is essentially a mesoscale feature embedded in the synoptic wind field, and that the diurnal cycle played an important role in both the wind maximum development and turning of the flow around the continental corner.

The synoptic evolution illustrated in Fig. 2b shows cyclonic flow at 850 hPa over the southeastern Australian region at 0000 UTC 10 November. The winds over the Bass Strait region are greater than 20 m s\(^{-1}\), are roughly oriented parallel to the coast, and are largely geostrophic (determined from the orientation of the geopotential height contours). East of Bass Strait the flow is largely westerly. Twelve hours later (Fig. 2c) the eastward migration of the anticyclone led to anticyclonic flow over the southeastern Australian region, and a significant southerly flow component over most of the Tasman Sea. During this period the flow offshore of Gabo Island shifted from a predominantly westerly direction to southwesterly, which illustrates the synoptic evolution that supported the turning of the wind maximum left into the Tasman Sea.

As mentioned in section 2, the Bass Strait wind maximum developed after the westerly synoptic flow was severely retarded by friction over the Victorian and Tasmanian landmasses. The 850-hPa geopotential height winds were relatively strong (20–22 m s\(^{-1}\), not shown) between 2000 UTC 9 November and 0000 UTC 10 November over a band centered on 40°S south latitude that extended across most of grid 1. However, they weakened during this period by about 10% and continued weakening thereafter as the high pressure system approached from the west. The north–south vertical section of wind speed, which spans Bass Strait (Fig. 6), demonstrates the effect of friction, generated by daytime convective mixing, on the wind speed. Funneling of flow between the Victorian and Tasmanian mountains (the latter not evident in Fig. 6) led to some flow acceleration through Bass Strait. The combined effect of topography and friction is the greatly reduced wind speeds over land to a depth of about 2000 m and the acceleration of flow in between, which produced a channel of relatively high wind speed through Bass Strait.

Two sensitivity experiments, one with the topography removed and the other with the topography multiplied everywhere by 1.5, confirmed that it was the deep friction layer over the land that was mostly responsible for the wind maximum and not funneling of flow between the Victorian and Tasmanian mountains. The wind speed in the simulation with enhanced topography was only 5%–10% greater than the simulation with no topography. Differences in wind speed over the land and Bass Strait were enhanced during the day when the convective mixing was most energetic.

The strong winds over Bass Strait and farther east over the Tasman Sea at 0000 UTC 10 November (Fig. 5a) were in near-geostrophic balance and had been for much of the previous 24 h. This near balance is illustrated in the Fig. 10a inset, which shows minimal ageostrophic forcing in that area. After 0000 UTC (1000 LST) the near balance was disrupted by continental heating that lowered the pressure over the continent, and the arrival of weaker pressure gradients with the approaching high pressure system (Fig. 10b). The ageostrophic forcing at 0400 UTC is illustrated in the Fig. 10b inset and shows particularly strong onshore forcing within 100 km of the coast from the coastal pressure gradient, and weaker northward forcing farther offshore from synoptic pressure changes. Both pressure changes led to flow acceleration to the left, which brought about the flow turning from a westerly direction through Bass Strait to southerly beyond the continent. The Fig. 10b inset shows that the forcing was strongest near the coast, and Fig. 5c shows that within the next 4 h the turning had led to onshore flow along the east coast south of Nowra.

As mentioned in section 2 advection of the cooler marine air with the onshore flow (Fig. 9b) led to the sudden pressure change at the coastal stations (Figs. 3d–f). The weaker forcing farther offshore led to more gradual turning with distance from the coast (cf. the flow direction indicated by the vectors in Figs. 5a–c). The reason for the forcing differences with distance
Fig. 9. Grid 2 potential temperature (contour interval 1 K) at 647 m for (a) 0400, (b) 0800, (c) 1200, and (d) 1600 UTC. The wind maxima shaded in Fig. 5 have been added to these figures to simplify visual comparisons between the two figures.

from the coast is evident from the pressure gradient implied from Fig. 10b. Immediately offshore of the coast between Gabo Island and Nowra the pressure gradient acts to the left of the wind, which is in the same direction as the Coriolis force (Southern Hemisphere). Farther offshore, the pressure gradient perpendicular to the wind becomes zero at the pressure maximum and changes direction beyond that so that it opposes the Coriolis force.

By 1200 UTC 10 November (2200 LST) the acceleration toward the continent in the vicinity of the coast had become negligible along the entire coast, except between Gabo Island and Nowra where the Coriolis force acting on the semistationary wind maximum (Fig. 5d) exceeded the opposing pressure gradient (not shown). Weaker ageostrophic forcing had continued to accompany the leading wind maximum, which ensured continued gradual turning to the left. The ageostrophic forcing in the vicinity of the leading wind maximum is depicted in the Fig. 10c inset. Note the closed contour of forcing greater than $4 \times 10^{-4}$ m s$^{-2}$ that roughly coincides with the wind maximum. Note also a region of northward-directed ageostrophic forcing of greater than $6 \times 10^{-4}$ m s$^{-2}$ caused by the unbalanced pressure gradient there.

Thus it would appear that the dominant mechanism responsible for initiating the flow turning around the continental corner after passing through Bass Strait is the continental heating that broke down the near-ageostrophic balance in the vicinity of the coast. On the
previous day when continental heating temporarily broke down the near-geostrophic balance, the flow turned to the left, but without the support of the broad-scale synoptic component it evolved into an anticyclonic eddy that propagated eastward out into the Tasman Sea (not shown). This shows that although the ageostrophic acceleration near the coast was considerably more intense, the broader scale and longer-lasting acceleration due to synoptic pressure changes was critical to the sustained flow turning, and the eventual intersection of the strongest part of the wind maximum near the Hunter Valley, which spawned the CTD north of the valley.

4. Discussion and summary

In section 2 a detailed analysis of the model winds and potential temperature distribution was performed and the coastal ridging explained. These results indicated that the ridging, which appeared to HL86 and RTJ99 to result primarily from various forms of CTD, could be explained by midlevel waves, the replacement of warm continental air with cool marine flow, and the blocking of this flow by the mountain barrier. On the south coast, where the mountain barrier to the north is particularly small (e.g., Mount Gambier and Laverton; Fig. 1) the more sudden pressure rise was explained primarily by the passage of a midlevel wave, and the onshore advection of a cooler air mass. Farther east blocking of the onshore flow also contributed to the ridging. On the east coast, the ridging was mostly the result of the onshore advection of cooler marine air (advected by the wind maximum that passed through Bass Strait and turned left over the Tasman Sea), which replaced a warmer continental air mass, and blocking of this flow against the mountain barrier.

The blocking is evident in Fig. 11 at 1200 UTC 10 November, which shows the difference in pressure between the original model simulation and a sensitivity
simulation in which the topography was removed. Note the positive values on the coastal side of the mountain barrier that result from pooling of denser air against the barrier by the persistent onshore flow, and the negative values on the opposite side of the mountain barrier where the denser flow had penetrated in the sensitivity simulation. This pooling of cooler air against the mountain barrier, with warmer air above, would, in theory, provide a suitable wave guide for CTD propagation.

In section 3 the development of the Bass Strait wind surge was explained, as well as the forcing that led to the turning of the wind once it entered the Tasman Sea, and the continued turning until it made landfall near the Hunter Valley. Surface heating played a crucial role in this process, and was supported by the broader-scale and more gradually changing synoptic evolution. Surface heating and the associated convective mixing retarded the flow over Victoria and Tasmania to a depth of about 2000 m, which led to the isolation of a zone of stronger winds through Bass Strait. The reduction in pressure over the continent led to strong ageostrophic forcing toward the coast. Weaker ageostrophic forcing developed farther offshore as pressure gradients weakened with the approach of the high pressure system from the west. The strongest ageostrophic forcing close to the coast and to the left of the flow direction, produced flow diffusence that led the wind surge on the coastal side, and an associated reduction in wind speed. Thus, the strongest component of the wind maximum remained farther offshore where it turned to the left more gradually and eventually made landfall near the Hunter Valley.

The resulting CTD that developed after this comparatively strong wind converged on the locally tall mountain barrier north of the Hunter Valley was short lived since the barrier beyond Coffs Harbour drops from about 1000 to 400 m within a couple of hundred kilometers. This highlights the relatively poor conditions for trapping in southeastern Australia compared with North America and partly explains the difficulty in identifying Australian CTD.

Although these model results failed to find a trapped disturbance south of the Hunter Valley, they do not disprove the existence of CTD in reality during this event or other events. However, given the good agreement between the model and observed pressure traces, the results do suggest that any trapped component accompanying a category 1 ridging event would have a secondary effect on the coastal pressure change. Further investigation into other southeastern Australian ridging events would be necessary to determine whether the 10–11 November 1982 event was an example of a typical non-SB category 1 ridging event, and to determine how common CTD are during these events.

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