Sea-breeze observations and modelling: a review

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We present an overview of sea-breeze characteristics and behaviour and then discuss some more important aspects in greater detail. These include the fine-scale structure of the sea-breeze in the head region, its behaviour in complex terrain, and its ability to penetrate hundreds of kilometres inland. These properties are then considered in a discussion of the role of sea-breezes in the dispersion of pollutants. Finally we examine the historical milestones of sea-breeze modelling, as it progressed from the early linear analytic models to the complex non-hydrostatic numerical models of the nineties.

Introduction

The sea-breeze has been studied extensively in the past, especially so since the end of World War II. Neumann (1973) refers to the Greek philosophers Aristotle, Plutarch and Theophrastus, and their ideas on the sea and land-breeze. They considered the sea-breeze to be a reflection from obstacles such as islands and coastal hills; onshore winds from the open ocean were not regarded as sea-breezes by them as there are no objects in the open sea from which the land-breeze could rebound. Jehn (1973) refers to Dampier (1697, 1701 and 1705) who wrote at length on winds, including land and sea-breezes. In his writings he gave a correct description of the sea-breeze which was ignored by the theoreticians for nearly 200 years.

Before the more detailed aspects of the sea-breeze can be investigated, it is necessary to describe the basic processes involved in its development. These processes are discussed in most of the elementary meteorology texts, the following being based on the account presented by Clarke (1955).

The sea and land-breeze circulations encountered near shorelines are due to the contrasting thermal responses of the land and water surfaces because of their different properties and energy balances. Because the sea is a fluid in continual motion, any heating or cooling is distributed to a considerable depth and so the rise or fall of water surface temperature is only slight. This results, for example, in a very small diurnal variation in surface temperature. Conversely, the land has a very small thermal conductivity and so heats and cools rapidly, resulting in a marked diurnal variation. The resulting land-water temperature differences and their diurnal reversal produce corresponding land-water pressure differences in the atmosphere which result in a system of breezes across the coast. Onshore flow during the day is known as a sea-breeze while offshore flow at night is referred to as a land-breeze.

During the day the heating of the near-land surface is distributed upwards by mixing. The rise in air temperature results in expansion, so that aloft, individual particles over the land are lifted relative to those over the sea. Level for level, the pressure aloft, but not at the surface, becomes higher over the land than over the sea, and this results in a compensating outflow of air from near-coastal areas over the land to those over the sea. As a result the surface pressure falls over a coastal strip of land and rises over a coastal strip of sea. The pressure gradient thus set up results in a low-level onshore flow with a return flow aloft. These factors operate simultaneously, not consecutively.

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The sea-breeze circulation cell begins near the coast in the morning and expands both landward and seaward. If there is no gradient wind the expansion is greater over the water due to a smaller frictional force at the sea surface than at the land surface. The effect of non-zero synoptic flow is to either accelerate or delay the development of the system. In some cases strong offshore gradient winds result in the sea-breeze developing over the sea with little or no inland progression during the day. In cases where the gradient wind is very light or has a moderate onshore component, the sea-breeze front tends to form 10 to 30 km inland in the afternoon.

The depth of the sea air, usually defined by the level of the zero onshore wind component, is often less than half the vertical extent of the sea-breeze system. At any given location the depth of the sea air varies with the time of day due to the migration of the system inland and to a thickening of the layer of sea air as the day progresses. On average the vertical extent of the sea-breeze is greater in tropical areas than in temperate zones. In temperate regions the sea-breeze inflow is usually between 200 and 500 m deep. These depths increase from 1000 m in moderately warm climates to 1400 m in tropical coastal regions. Maximum heights of 2 km have been observed in India (Defant 1951; Atkinson 1981).

The return flow is often difficult to observe as it is often 'masked' by the synoptic conditions. Lyons (1972) found that the return flow layer of Lake Michigan breezes was typically twice as deep and with peak velocities half those of the inflow layer. Atkinson (1981) summarises observations which show the depth of the sea-breeze return flow ranging from less than 0.5 to nearly 3 km.

The sea-breeze is usually accompanied by a decrease in the temperature and a corresponding increase in the humidity of the atmosphere. Atkinson (1981) presents data for the sea-breeze in India, showing a mean monthly temperature fall due to the sea-breeze varying from 1.1°C to 4.4°C. These same data show that the range of relative humidity rises is from five to 30 per cent with an average value of about ten per cent. The observations of Simpson et al. (1977) show that the main humidity rise takes place within 200–250 m of the leading edge of the front. They also found a minimum value of humidity directly behind the front, indicating dry air which has descended behind the front and which will subsequently be mixed with the moist sea air. This minimum most often occurs about 800 m from the front. The transport of moister air over the land is regarded as a typical characteristic of the sea-breeze, although this is by no means definitive. Observations presented by Physick (1982) and Abbs (1986) document cases in which the reverse was observed to occur.

Flohn (1969) gives examples of the average horizontal extent of the sea-breeze for various climatic regimes. In mid-latitudes the horizontal extent of the sea-breeze rarely exceeds 40 to 50 km, while in tropical and subtropical areas it reaches 100 to 150 km on both the landward and seaward sides of the coast, resulting in a total extent of 200 to 300 km. Some of the most extensive studies of deeply penetrating sea-breezes have been carried out in Australia by Clarke (1955) and Reid (1957). Clarke has recorded cases of the sea-breeze reaching Kalgoorlie, 300 km from the coast. Clarke (1983a) also reports results which indicate that the sea-breeze penetrates 500 km inland in northern Australia and has also been recorded 320 km seaward of Borneo.

In the following sections, we examine some aspects of the sea-breeze in greater detail, including its deep inland penetration and its role in determining air quality in coastal regions. We conclude with a summary of the more important milestones as sea-breeze modelling progressed from the early linear analytic models to the complex non-hydrostatic numerical models of the nineties.

Fine-scale structure of sea-breezes

The similarities between density currents produced in the laboratory, and atmospheric examples such as sea-breezes and thunderstorm outflows, were first shown at Reading University by retired school master John Simpson. Denser (sea) air flows towards the sea-breeze front at low levels and is swept up and backwards at the leading edge in the region referred to as the head (Simpson 1969, 1972). Mixing between cool moist sea air and warm dry land air takes place at the rear of the head through breaking billows, generated at the front of the head by the Kelvin-Helmholtz shear instability mechanism. Lidar observations of such billows, with a wavelength of about 2 km, can be found in Nakane and Sasano (1986). The mixed air flows back towards the coastline above the inflow and constitutes the return flow of the sea-breeze.

Later experiments by Simpson and Britter (1979) examined the relation between the depths of the boundary layer, the inflow and the raised head. Nakane and Sasano (1986) measured a 1300 m head with a following flow of 300 m, but no mixed-layer height ahead of the sea-breeze is given (the Simpson and Britter experiments imply it should be about 5000 m, which is surely too large). For the well-documented sea-breeze described by Simpson et al. (1977), also with a 300 m deep inflow and with a mixed layer of 1400 m, the head was found to be only twice the depth of the 300 m deep inflow and extended back about one to two kilometres from the leading edge. Aircraft measurements through this sea-

Fig. 1 Plot of vertical velocity, humidity mixing ratio and potential temperature made during a traverse of a sea-breeze front. The traverse was made at a height of 600 m. The sea-breeze front occurs at 0 km on the horizontal axis. The captions 'rough', 'bubbly' and 'steady, smooth' indicate the degree of turbulence (Simpson et al. 1977).
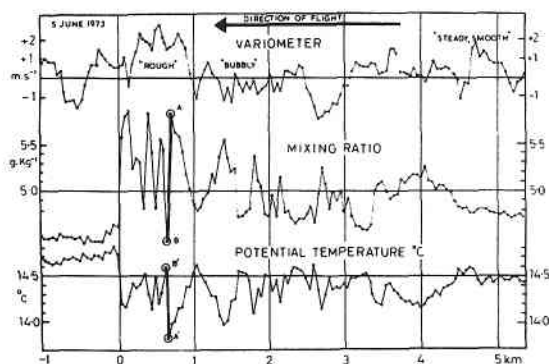
breeze (Fig. 1) show mixing taking place in the head region with a billow wavelength of about 200 m. Points marked A and B, 50 m apart, denote adjacent data points characteristic of undiluted sea and land air respectively.

Along the front during the day, the leading edge consists of a continually changing pattern of lobes and clefts about one kilometre across. These arise from convective instability caused by the overrunning of warm land air at the ground by denser sea air (Britter and Simpson 1978), and disappear at night when radiative cooling rapidly cools the near-surface layers.

In a summary paper of the laboratory experiments on density currents, Simpson and Britter (1980) point out that no mixing takes place at the interface between the sea-breeze inflow and the return flow above. This implies that emissions of pollutants into the sea-breeze will be transported inland virtually undiluted, rise and mix with ambient air at the front, and head back towards the coast in the return flow. Furthermore, a 'clean' sea-breeze will sweep polluted ambient air upwards and remove it seawards at higher levels while maintaining unpolluted air at the ground. The role of sea-breezes in air quality is discussed in a later section.

A recent aircraft study of sea-breezes on the straight South Australian coastline by Kraus et al. (1990) obtained data at 3 m resolution at various levels through the fronts. The various terms in the frontogenesis equation were calculated on a 2 km length scale, with confluence, shear and diabatic processes contributing in equal magnitude to

frontogenesis. The raised head in this study is about twice as deep as the 250 m deep sea-breeze inflow.

The first fine-resolution modelling study of the sea-breeze to appear in the literature was published by Sha et al. (1991). Using a non-hydrostatic model and a horizontal resolution of 100 m, they were able to simulate all the fine-scale features found in the laboratory experiments and verify the empirical relations involving the inflow and head depths, and billow amplitudes and wavelengths. They also found that Kelvin-Helmholtz instability (KHI) does not occur in the initial or final stages of the sea-breeze, and that the KHI-induced mixing in the middle stage (afternoon) is responsible for slowing the inland penetration during this period. This finding may explain the failure of coarser-grid models to simulate the observed middle-of-the-day slowing of the 14 June 1973 sea-breeze (Carpenter 1979; Physick 1980).

Behaviour in complex terrain

The topography of the coastline has an important controlling effect on the sea-breeze. As well as influencing the inland penetration of the sea-breeze it affects convergence/divergence which in turn affects associated weather. It may also affect the number of sea-breezes experienced and there are indications that it may have an effect on the onset time. The classical model of the sea-breeze, as outlined in the introduction, has been developed in uncomplicated terrain, usually along a long straight coastline with a flat hinterland. Later studies conducted in complex terrain indicate that these sea-breezes do not conform to the simple model.

Irregular coastlines

Curved coastlines. Neumann (1951) showed that the convergence or divergence of the land and sea-breezes depends on the curvature of the coastline. Where the coast is concave seawards the sea-breeze diverges in contrast to the converging land-breeze, while at convex coastlines the reverse occurs. He also showed that the variation in the percentage of thunderstorms within coastal areas is connected in part with the convergent or divergent nature of the sea and land-breezes involved.

In a study of the meteorological conditions associated with Los Angeles air pollution, Edinger and Helvey (1961) discovered that on some summer afternoons, narrow zones of converging winds extend across the inland valleys. The continuous records from a group of observing stations in the San Fernando valley revealed abrupt shifts

from the usual sea-breeze from the south-east to a westerly sea-breeze later in the day. These converging winds are the result of the sea-breezes formed along a convex section of the California coast. As the sea-breezes converge, both easterlies and westerlies are deflected upwards in the convergence zone and become strong updrafts. They suggest that this mechanism may help to act as a generator of upper level pollution by pumping aloft Los Angeles air pollution. Similarly, in the Elsinore Valley, glider pilots have reported the occurrence of a 'smog front' on many summer afternoons. This results from the convergence of two air masses of markedly different histories over the land – one a polluted breeze from Los Angeles, the other an unpolluted breeze from the coast. They also report that the line along which the two sea-breezes converge is the site of strong vertical currents.

The convergence of sea-breezes from different coastlines has also been studied by Physick and Abbs (1991, 1992). Their analysis of data from the Latrobe Valley surface network (Fig. 2) revealed that sea-breezes from the south and east coasts of the region were the dominant wind features over the region during the study period. In this case the east coast sea-breeze travels farther inland and is the stronger of the two breezes, pushing the southern sea-breeze back towards the coast after the two meet in the late afternoon. This behaviour was reproduced in their numerical modelling study.

The first three-dimensional modelling study of the sea-breeze (McPherson 1970) was motivated by a desire to assess quantitatively the effect of coastline curvature on the evolution of the breeze. In his simulations he was able to show that the effect of the bay is evident immediately. Both the sea-breeze convergence zone and the thermal gradient were distorted around the model bay and were symmetric with respect to the bay. As the simulation progressed the degree of distortion diminished and an asymmetry in the thermal and vertical motion fields developed. This asymmetry is due to the Coriolis acceleration.

Bay and ocean breezes. In regions where the coastline changes direction suddenly, such as a gulf or deep bay, it is possible to produce two sea-breezes. The first of these we will refer to as the bay-breeze. This breeze is a response to the temperature difference between the waters of the bay (or gulf) and the surrounding coastline. The second is the ocean-breeze which originates along the 'large scale' coastline.

Physick and Byron-Scott (1977) observed the existence of sea-breezes on each side of St Vincent's Gulf and an ocean-breeze which moves up the Gulf from the Southern Ocean and reaches Adelaide during the afternoon. The ocean-breeze is evident in the wind records only under near-calm synoptic conditions. This is supported by

Fig. 2   Mean arrival time (in the form of isochrones) throughout the Latrobe Valley for the period 8–12 March 1985 for (a) the sea-breeze from the east coast and (b) the sea-breeze from the south coast (Physick and Abbs 1992).
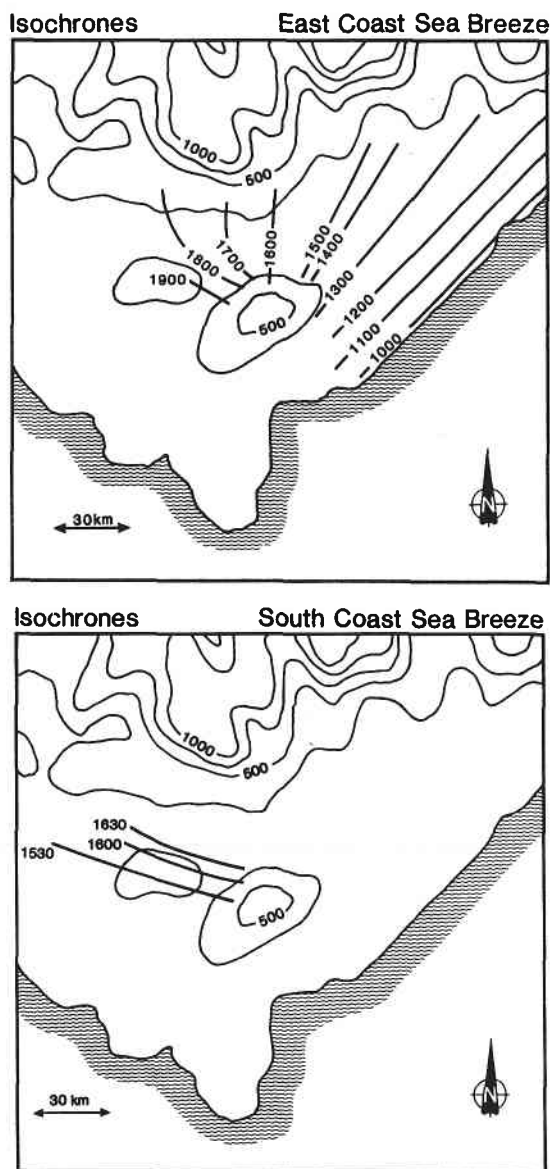
examples of days when the gradient wind is east to northeast, the resolved wind at the western coastline is observed to be from the southeast in the afternoon whereas the Gulf sea-breeze component and the gradient wind component are both from the northeast quadrant.

Barbato (1978) observed the generation of a bay-breeze and an ocean-breeze in Boston. Boston is situated in a concave, hemispheric-shaped basin with an escarpment rising to 100 m located along the west and northwest. He found that the bay-breeze was often superseded later in the day by the ocean-breeze. Ocean-breezes of long duration penetrated inland with well defined characteristics.

A similar situation has been observed by Abbs (1986) for Port Phillip Bay, Melbourne. She observed situations in which the ocean-breeze from Bass Strait penetrated inland above the Port Phillip Bay bay-breeze but was not recorded at the surface. Under these conditions the ocean-breeze ascends the bay-breeze and penetrates inland with it. This situation occurs when the ocean-breeze is warmed from below by the land surface over which it passes, hence becoming more buoyant. Because of this excess buoyancy the ocean-breeze is able to mount the bay-breeze as it moves inland. On many occasions when the ocean-breeze does occur at the surface, its onset is accompanied by an increase in dry-bulb temperature and a decrease in humidity. This is due to the advection of warmer, drier, land-modified air replacing the cool, moist bay-breeze of Port Phillip Bay. It has also been shown, both observationally and numerically, that the interaction of these two breezes may produce a mesoscale cyclonic circulation.

Orography

Does the presence of coastal orography act as a barrier to the inland penetration of the sea-breeze? Numerical experiments by Mahrer and Pielke (1977) using a 900 m high bell-shaped mountain extending from the coast to about 75 km inland found that the sea-breeze and mountain circulations acting together produce a more intense circulation than they do acting separately, and that by early afternoon the sea-breeze had reached the lee side of the mountains. This general result was also found for the Kanto Plain region of Japan by Kikuchi et al. (1981) who showed that the extended sea-breeze (over 200 km horizontally) observed in that region could only be produced numerically when the orography was included. Similarly, strong afternoon winds regularly observed over Lake Kinneret in Israel were shown by Alpert et al. (1982) to be the sea-breeze from the Mediterranean coast 45 km away. During its journey to the lake (210 m below sea level), the sea-breeze passed over hills 400 m high. In their numerical experiments for the Perth region of Western Australia, Manins et al. (1992) similarly found that the presence of the escarpment, 30 km from and parallel to the coast, acts to accelerate the inland movement of the sea-breeze.

Observations from coastal valleys running parallel to the shoreline, and opening to the sea at some location, have shown that the sea-breeze will either travel up the valley from the mouth or spill over the coastal ridge into the valley, depending on the direction of the synoptic wind (Sumner 1977; Physick 1982). Studies on the west coast of the United States have shown that the inland penetration of marine air is strongly dependent on topographic channels (Fosberg and Schroeder 1966; Olsson et al. 1973). In the San Francisco region, Fosberg and Schroeder (1966) have found that the sea-breeze moves from the coast to the inland valley through the natural passes such as the Golden Gate, Petaluma Gap and Carquinez Strait. They also found that these passes and deflecting barriers such as hills create vortices in the sea-breeze flow. During the Atmospheric Studies in Complex Terrain (ASCOT) program in the Geysers area of northern California, the sea-breeze was found to have a strong influence on the initiation and evolution of evening drainage winds (Neff and King 1987).

Neumann and Savijarvi (1986) modelled the situation where mountains rise steeply from the shoreline. They were prompted by two sets of 19th century sea-breeze observations from the mountainous coasts of Chile and Peru, involving sailing ships. In one of these, the vessel *Antuco* lay in port ready to sail for four days 'while the water about her was mirror-smooth'. About 4–6 km offshore, the sea was disturbed each day by a moderate sea-breeze. In their experiments they varied the mountain height and the Brunt-Väisälä frequency, finding that as each is increased the blocking effect on the sea-breeze intensifies. This effect is most pronounced in the vicinity of the shoreline, where the sea-breeze breaks down into two cells (Fig. 3), one over the sea and another

Fig. 3 Isotachs of the onshore component (solid lines, in m s^{-1}) and of the vertical velocity (dashed lines, in cm s^{-1}) of the sea-breeze flow in the presence of a 1000 m high barrier. Note that in the detached cell over the lower slope of the mountain the maximum seaward velocity is -3 m s^{-1} and the maximum downward velocity is -28 cm s^{-1} (Neumann and Savijarvi 1986).
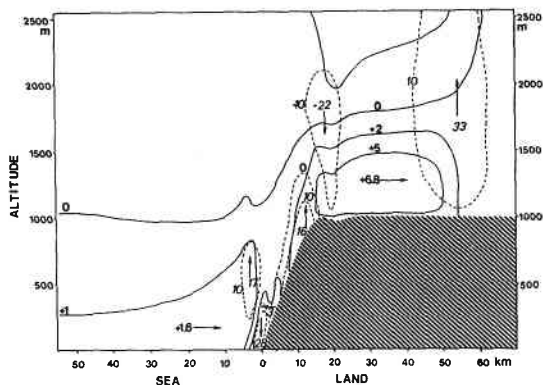
over the windward side of the mountain. For the sailing ship case, their predictions were in agreement with the observations of the ship's captain.

Inland penetration

While working as a forecaster in Canberra in the 1940s, Reg Clarke became interested in a late afternoon easterly wind which was often observed in Canberra during the warmer months (Clarke, personal communication). Certain that it was a sea-breeze from the east coast (112 km away) but unable to convince any of his colleagues, Reg used car and aeroplane observations to prove that this was indeed the case. Data from these experiments, in which it was also claimed the sea-breeze progressed to Wagga Wagga (270 km inland), were published much later and favourably compared to Reg's numerical predictions of the inland penetration (Clarke 1983a).

To avoid any orographic effects, and interested in the claim of Hounam (1945) that sea-breezes occasionally reached Kalgoorlie (345 km inland), he later mounted a sea-breeze experiment in the featureless terrain of southern Western Australia (Clarke 1955). To prove his assertion that sea-breezes were not just coastal phenomena (the prevailing view at the time), Reg and his brother tracked several sea-breezes by car from the south coast through to Kalgoorlie, where they arrived around midnight. Since this pioneering work, sea-breezes have been observed by Reid (1957) at Renmark (217 km inland), by Garratt and Physick (1985) at Daly Waters (280 km) and by Clarke et al. (1981) across Cape York Peninsula. Elsewhere, deeply penetrating sea-breezes have been observed up to 200 km from the coast in California (Carroll and Basket 1979), in Japan (Kurita et al. 1985), and in Saudi Arabia (Steedman and Ashour 1976).

One of the most interesting aspects of the penetration of the sea-breeze is the change in the rate of advance inland of the front, although conflicting reports on this characteristic appear in the literature. Sha et al. (1991) identified three distinct stages in the penetration of their modelled sea-breeze front. Initially the front moves at a speed of approximately 5 km h^{-1} before decelerating in the early afternoon. Later, the sea-breeze front accelerates before decelerating and decaying late at night. The afternoon deceleration of the sea-breeze was observed by Pedgley (1958) for sea-breezes in Egypt and by Simpson et al. (1977) for their well documented southern England sea-breeze. The evening acceleration of the sea-breeze front has been observed and modelled by a number of authors (Clarke 1955; Simpson et al. 1977; Clarke 1984; Physick and Smith 1985).

Simpson et al. (1977) were able to show, using a simple numerical model, that the inland advance of the sea-breeze is governed locally by the density differences across the front. Under conditions in which the synoptic flow across the coastline is weak, a sea-breeze will develop at the coast as soon as the land-sea temperature difference is large enough. As the sea air moves inland its temperature rises more rapidly than that of the land air as the heat is distributed over a much smaller depth. Consequently, the temperature contrast across the front decreases and the sea-breeze front decelerates. In addition Kelvin-Helmholtz instability increases the top drag through strong turbulent mixing and this increase in the top friction is also associated with the deceleration of the sea-breeze in the middle of the day (Sha et al. 1991). In the late afternoon the solar heating decreases but the temperature of the land air still increases gradually. At the same time, the temperature of the sea air is becoming cooler as the sea air arriving at the front has had less heat added to it since crossing the coast. As a result, the temperature difference across the front increases and the sea-breeze begins to accelerate. In the nocturnal stage of the sea-breeze, horizontal advection of cool sea air is instrumental in maintaining the frontal density gradient. This advection of cool sea air ceases once air which crossed the coast near sunset reaches the front. Physick and Smith (1985) suggest that the further inland a sea-breeze is at sunset, the longer it will survive before dissipating in the manner described below.

Clarke (1984) used a two-dimensional hydrostatic numerical model to investigate further the behaviour of sea-breezes when the geostrophic wind is onshore. He identified five stages in the development of the sea-breeze. During the 'immature stage' the inland penetration increases monotonically with time. At this time the relative flow is through the isentropes and the flow is far from steady state. Convective overturning maintains the frontal boundary in a near-vertical position. In the 'early mature stage', corresponding to late afternoon, the sea-breeze surge is still not a steady-state gravity current but acceleration of the surge has commenced. During early evening the supply of cool air from the sea ceases, decoupling of the frictional link with the surface occurs and the sea-breeze continues to accelerate. Clarke identifies this as the 'late mature stage' of the development of the sea-breeze surge. At this time the leading isentropes are still steeply inclined but are beginning to be flattened and the centre of the horizontal vortex shifts to the leading edge of the circulation. The surge front is still sharp and active during this stage. In the next stage, the 'early degenerate stage', the layer of cold air flattens and spreads inland propagating as an unsteady gravity current. During this period the sea-breeze steadily decelerates and has the form of a degenerate gravity current or a steadily diminishing vortex com-

pletely detached from the coastline. Figure 4 shows a vortex about 10 km in length observed during the last stages of a sea-breeze surge. In this vortex the air is moving faster than the front. Physick and Smith (1985) showed (using a two-dimensional model) that once the vortex has formed, the main heating/cooling process that occurs is the subsidence warming in the descending arm of the circulation. This leads to the eventual decay of the density gradient and the vortex itself. In the late degenerate stage there is no longer a closed circulation near the leading edge.

Fig. 4 Relative streamlines for a dying sea-breeze surge observed during the Coonalpyn Downs expedition in South Australia. The streamlines are determined from serial balloon flights centred on 2200 CST at a point 165 km inland (Clarke 1983a).
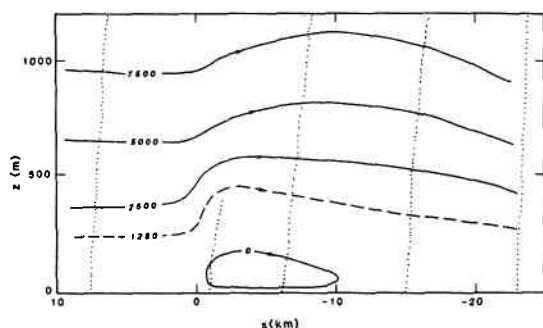
## Role in internal bore generation

The 'morning glory' of northern Australia is the most widely documented example of internal undular bores occurring in the atmosphere, and the importance of the sea-breeze in its formation has been accepted since the early 1980s. A complete description of the 'morning glory' is presented in this volume by Christie (1992). Analysis of the field data from the 1978 and 1979 expeditions, enabled Clarke et al. (1981) to show that the morning glory is not a gravity current, a characteristic of which is low-level feeder flow toward the leading edge of the disturbance in a frame of reference moving with the front. Their observations showed that the relative flow was through the system and contained very little advected fluid. The streamline patterns show that the disturbance has the characteristics of a travelling wave (or group of waves) but the pressure rise accompanying the disturbance is sustained for over three hours. Thus, they conclude that the disturbance is an atmospheric undular bore propagating on the nocturnal or maritime inversion.

Clarke et al. (1981) present four possible mechanisms for the production of the pressure jump associated with the morning glory. Their data suggested that the dominant mechanism for formation of the pressure jumps was the interaction of the east coast sea-breeze gravity current with the nocturnal inversion. The sea-breeze develops a front as it progresses inland assisted by the prevailing light onshore geostrophic flow. With time, the front may develop into a closed vortex which may persist for many hours following sunset. This explanation was supported by observations of deeply penetrating sea-breezes (Clarke 1965). The numerical modelling studies of Crook and Miller (1985) and Sha et al. (1991) confirm that it is possible for an undular bore to be produced by a gravity current moving into a stably stratified fluid. Other suggested mechanisms were the production of an internal bore on a katabatic flow, the collapse of lee waves or standing eddies in the vicinity of high ground and the head-on collision of sea-breezes from the east and west coast of Cape York Peninsula.

After further analysis of the field data, Clarke (1983b) concluded that the morning glory is produced by the interaction of sea-breezes from both sides of the Peninsula. The first necessary condition for bore formation is that a sea-breeze surge should move inland from the east coast of Cape York Peninsula and interact with the weaker sea-breeze on the west coast of the Peninsula. The west coast sea-breeze acts to modify the atmosphere in a coastal strip about 100 km wide, producing a nocturnal boundary layer over both sea and land on which the internal bores can propagate. For this sequence of events to occur a modest easterly geostrophic wind is necessary.

This description of the formation of the internal bore was later confirmed by the two-dimensional modelling studies of Clarke (1984) and Noonan and Smith (1986). Clarke showed that the collision of the two sea-breezes from either side of the Peninsula raises cool air in a sharp, upward bulge. One or two bores may then propagate away from the genesis site at the same speed as the original sea-breezes. The evolution of the vertical velocity and potential temperature fields associated with morning glory genesis are shown in Fig. 5. Noonan and Smith also showed that sea-breeze collision must occur in the late evening for a sustained bore to be formed. This ensures that air associated with the east coast sea-breeze has stabilised at low levels and so there is a local upward flux of negative buoyancy on collision. If collision takes place earlier, the low-level air of the east coast sea-breeze is still well mixed and it is this air that is lifted at collision. Under these conditions a sustained bore is not produced. The three-dimensional studies of Noonan and Smith (1987) showed that the sea-breeze convergence is maximised in the late evening at about the time and place morning glories are observed to

Fig. 5   Vertical cross-sections of the vertical velocity and potential temperature (a) before, (b) at the time of, and (c) after morning glory genesis. In (d) the bore-like disturbance has begun to move westwards (Noonan and Smith 1986).
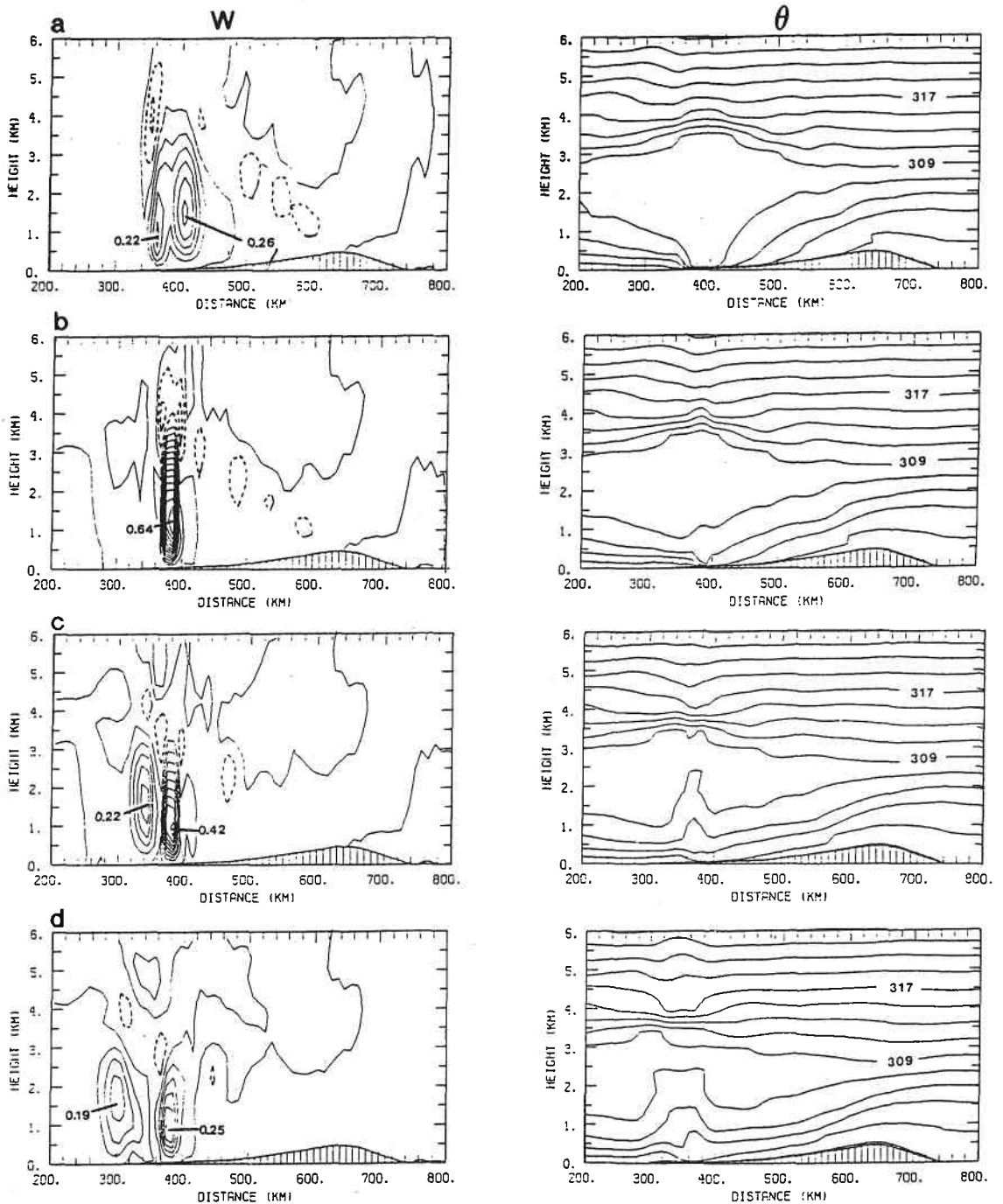
originate and a sustained westward propagating disturbance is produced. The disturbance is approximately parallel with the east coast of the Peninsula, reflecting the orientation of the east coast sea-breeze front. They conclude that the reason that morning glories are confined to the southern part of the Peninsula is due to the fact that by late evening the disturbance produced by the colliding sea-breezes has significant amplitude only in that region.

Sea-breezes and air quality

When assessing the impact of pollutant emitters in coastal regions, it is essential to consider several of the sea-breeze characteristics discussed in earlier sections. We examine the air quality in both the near and far field.

Near-source dispersion

The thermal internal boundary layer (TIBL) which develops in the stable sea air as the sea-breeze passes over the heated land can be responsible for enhanced surface concentrations downwind of a coastal source. Figure 6 illustrates how a plume initially travelling in stable air above the TIBL is brought to ground (fumigated) further inland by convective eddies in the growing TIBL. Probably the most comprehensive data set on TIBL behaviour and plume fumigation was gathered in 1978 at Nanticoke on the shore of Lake Erie, Canada (Portelli 1982; Kerman et al. 1982; Hoff et al. 1982). Over the eight days of the experiment, fumigation of the plume from the coal-fired power plant (final-rise height about 350–400 m above the ground) usually began between 1100 and 1300 hours local time and continued till about 1700–1900 hours. The plume was first brought to ground at distances ranging from 7 to 20 km from the stack, but this distance varied during the day as the TIBL grew and decayed. The fumigation zone was also found to rotate about the stack as the wind direction changed, primarily due to the Coriolis force acting on the sea-breeze.

Continuous fumigation of shoreline plumes was also examined by Lyons and Cole (1973) for the Milwaukee area who found that while the onshore lake-breeze was clean at the shoreline, elevated plumes at the shore were brought to ground in the TIBL within 3–5 km inland. A similar study by Lyons and Olsson (1973) on the western shore of Lake Michigan in the Chicago area found that not only does the same fumigation mechanism occur there, but the air quality problem is exacerbated by recirculation of pollutants within the lake-breeze circulation. Pollutants fumigated into the inflow rise at the front and move back over the lake in the return flow. However the larger pollutant particles subside into the top region of the inflow and are brought back over land where they are fumigated to the ground once again, along

Fig. 6 Schematic diagram illustrating fumigation of a coastal plume by the thermal internal boundary layer. This diagram is from Portelli (1982) and illustrates the data network for the Lake Erie fumigation experiment.
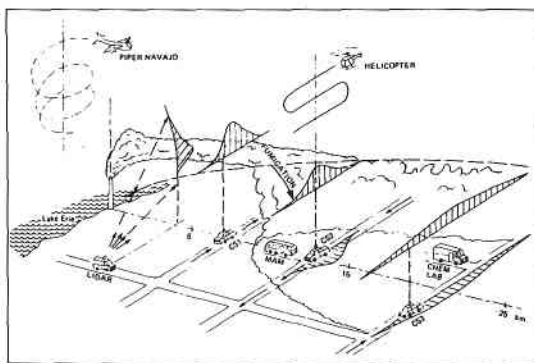
with the fresh emissions from the coastal sources, leading to an ever-increasing loading of the atmosphere in the shoreline region.

Our discussion so far has concerned elevated sources, primarily emitting oxides of sulfur and nitrogen. We now turn to surface emissions, of reactive pollutants, and show that another aspect of the sea-breeze is frequently responsible for high levels of photochemical smog, of which ozone is the major component.

Ozone is formed on a time-scale of several hours through reactions involving non-methane hydrocarbons and nitrogen oxides in the presence of ultraviolet radiation. The dominant source of these ozone precursors in populated areas is the internal combustion engine, although various industries also contribute. In coastal cities such as Athens (Lalas et al. 1987; Gusten et al. 1988), Melbourne (Evans et al. 1981; Cope et al. 1990; Johnson and Cope 1990), Sydney (Hyde et al. 1978; Hawke et al. 1983) and Perth (Manins et al. 1992), the local meteorology in summertime is such that land-breezes or light offshore synoptic winds advect emissions from the morning peak traffic period out over the sea. Here the mixture is able to 'cook' for a few hours unhindered by sources of fresh nitric oxide (NO) which are likely to reduce the ozone levels. When the sea-breeze develops in the late morning, the ozone is transported onshore and over the city and suburbs responsible for its precursors. Under these conditions, the highest ozone levels (in late afternoon) are usually found within 30–40 km of the coastal source. In the following section on long-range transport, we find that daily maximum readings of 'city-produced' ozone can occur as far away as 200 km from the source.

Long-range transport

The deep inland penetration achieved by some sea-breezes suggests that pollution from coastal regions may be transported large distances from the source. Such evidence has been found in a number of Californian localities including Yosemite National Park where the sea-breeze and upslope circulation combine to import pollutants from the San Francisco Bay area 200 km away (Carroll and Baskett 1979), and in Palm Springs where Los Angeles smog arrives from 180 km away around midnight (Grosjean 1983). Los Angeles is also responsible at times for high ozone readings in the Santa Barbera region to the north-west (Moore et al. 1991; Hanna et al. 1991).

In Japan, the highest ozone concentrations are not usually found in Tokyo where the precursors are produced, but in the mountainous region 150 km to the northwest in the early evening hours. The smog is transported there by a circulation resulting from interaction between the sea-breeze and thermally and topographically-induced flow, known as the extended sea-breeze (Kurita et al. 1985; Chang et al. 1989; Kondo 1990). By analysing meteorological and air quality data for the region and aircraft data along the route of a polluted air mass from Tokyo, Ueda et al. (1988) found that the highest surface concentrations of oxidants occurred at night-time in the mountains, at the rear edge of the cut-off vortex where the internal circulation brings upper-layer oxidants to the ground. This vortex, about 25 km in length in the Tokyo cases, forms at the leading edge of the sea-breeze as it encounters a stable atmosphere after sunset.

So far our discussion has concentrated on cases in which the sea-breeze exacerbates pollution problems. We now examine a situation in which it is very effective in replacing polluted surface air with clean air. The Latrobe Valley in southeastern Australia contains coal-fired power stations located about 90 km up the valley from the coast. Under clear-sky summertime conditions, the station plumes become well mixed throughout the convective layer and drift down the valley, meeting the sea-breeze about 25 km from the coast. Air in the polluted mixed layer rises at the sea-breeze front, is mixed with incoming sea air at the top and rear of the head region and continues its passage towards the coast in the return flow at higher levels. Clean air is brought in at lower levels as the sea-breeze continues its movement inland. By the time it reaches the power stations in late afternoon, convection has decayed and the plumes are no longer brought to ground. They are taken further inland during the night at about 500 m above the ground, while at higher levels towards the coast the sea-breeze return flow transports afternoon emissions out to sea. For the Latrobe Valley region, it appears that the emitters may very well be located at the optimum location, as

far as minimising air pollution on sea-breeze days is concerned. Further discussion on the meteorology and air quality of the Latrobe Valley region can be found in Physick and Abbs (1991, 1992).

Sea-breeze modelling

Whilst the dynamics of the sea-breeze must have occupied the minds of many natural philosophers during the last few hundred years, one of the earliest worthwhile theoretical studies of the phenomenon appears to be that of Jeffreys (1922), who considered it to be a type of antitriptic wind. Since then, a number of linear models of the sea-breeze have been produced, including that of Haurwitz (1947). He showed that the incorporation of friction into the model enabled the theoretical results to resemble the observations more closely, insofar as it caused the sea-breeze to decrease earlier. He also demonstrated that the observed veering of the sea-breeze during the day is due to the Coriolis force. Schmidt (1947) developed a linear model in which the velocity fields were calculated from an atmospheric temperature distribution which was specified as a function of time.

The linear models reproduced certain observed features of the sea-breeze, such as the depth of the flow and the rotation of the wind vector. However, it was not until Pearce (1955) performed the first successful numerical integration of the nonlinear sea-breeze equations, that models were constructed which exhibited the oft-observed sea-breeze front or convergence zone.

In a mesoscale phenomenon such as the sea or lake-breeze, there is a significant transport of momentum and heat by turbulence, and thus it is important that this aspect of the numerical model be simulated as accurately as possible. Pearce (1955) prescribed an artificial heating mechanism of the atmosphere unrelated to the predicted temperature and wind fields, and also neglected surface drag. In one of the first attempts at a more realistic simulation of the momentum and heat exchange, Fisher (1961) used a gradient transfer approach and assumed a constant profile for the eddy exchange coefficients (K_M and K_H) throughout the day and night, with a maximum value occurring at a height of 50 m. It is well known, however, that these transfer coefficients are at least an order of magnitude less in stable than in unstable conditions.

Estoque (1961) divided the modelled region into a constant-flux surface layer of 50 m, in which K_M ($=K_H$) is a function of stability and wind shear, and a transition sub-layer, in which K_M decreases in a linear fashion from its value at the

interface to zero at the top of the model at 2 km. This model represents a landmark in sea-breeze modelling and all advances since then, mainly in parametrisation of the boundary layer, have really been incremental. McPherson (1970) showed that the manner in which K_M decreases to zero at the top of the model is a significant factor in the determination of the intensity and extent of the computed sea-breeze circulation. An exponential decrease was considered by McPherson to give results in reasonable agreement with sea-breeze observations.

The next step forward followed the significant advances made in boundary-layer meteorology in the fifties and sixties. In the numerical model developed by Pielke (1974) to study sea-breezes in Florida, the non-dimensional wind and temperature profiles based on Monin-Obukhov similarity theory were used to calculate fluxes in the surface layer. The heights of the surface layer and boundary layer were also allowed to vary over a diurnal cycle. Surface temperature (T_0) had always been specified as a sinusoidal function of time in sea-breeze models to this stage, but Physick (1976) employed a surface energy balance equation to calculate T_0, thus allowing horizontal gradients of T_0 and surface heat flux to exist across the front. Development of this model benefited from discussions with Reg Clarke who was designing a model to help explain his observations of sea-breezes up to 300 km from the coastline. Similarity theory and a local scheme for the exchange coefficients throughout the boundary layer, developed from analysis of the Wangara data (Clarke 1974), were used in his model but details were not published until a decade later in a study of nocturnal wind surges (Clarke 1983a). However some results on deeply penetrating sea-breezes were presented at a Fire Weather Conference at Monash University (Clarke 1973).

Monin-Obukhov similarity theory in the surface layer and an energy balance equation for the computation of surface temperature are now standard fare in models used for the simulation of sea-breezes. The main variation between models occurs in the parametrisation of turbulent exchange in the transition layer above the surface layer. These range from local (Kondo 1990) and non-local (Pielke 1974) K-theory to first-order (Arritt 1987), second-order (Schumann et al. 1987) and third-order (Briere 1987) turbulent kinetic energy (TKE) schemes. A comparative study of a number of these parametrisations by Mahfouf et al. (1987) has found that the mean fields of wind and temperature are relatively insensitive to the turbulence specification, but that the more detailed schemes predict the turbulence statistics more accurately, an important finding for air quality models in which the dispersion is driven by winds and turbulence fields from a mesoscale model.

Virtually all mesoscale models used for sea-breeze simulations employ the hydrostatic assumption in the vertical equation of motion and use a horizontal grid spacing ranging from 2 to 10 km. Pielke (1972), Martin and Pielke (1983) and Song et al. (1985) have all examined the effect of the hydrostatic assumption in sea-breeze models by deriving an estimate for the non-hydrostatic pressure, based purely on results from a hydrostatic simulation. As well as concluding that the hydrostatic assumption is adequate for grid spacings as small as 1 km, they also investigated the dependence of the difference on the scale of the heating, atmospheric stability and synoptic wind speed, finding that non-hydrostatic effects generally tend to weaken the sea-breeze when compared to a hydrostatic simulation. Their conclusions were corroborated by Yang (1991) who compared hydrostatic and non-hydrostatic simulations from the same model. Yang also stressed the importance of the turbulence parametrisation and recommended that higher order closure schemes should be used in non-hydrostatic models.

Perhaps the final stage in sea-breeze modelling was reached recently with the publication of a paper by Sha et al. (1991). Using a non-hydrostatic model and a horizontal grid spacing of only 100 m, they were able to resolve explicitly the mixing process in the head region, whereby warmer ambient air is entrained into cooler sea air by the breaking of Kelvin-Helmholtz billows. This fine structure of the sea-breeze head as modelled by Sha et al. is shown in Fig. 7.

Fig. 7   **Vertical cross-section showing the fine structure of the sea-breeze head. The numbers on the horizontal axis are distance from the coastline. The solid line indicates the zero velocity boundary. The front position is also shown. Horizontal vortices are labelled, a, b, c (Sha et al. 1991).**
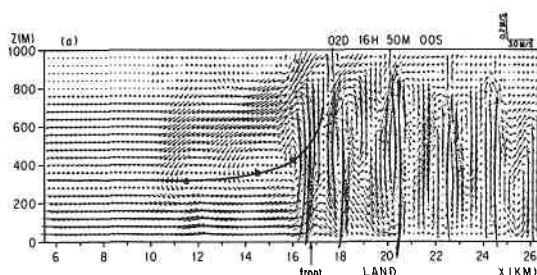
Conclusion

In this review we have presented early results which described the structure of the sea-breeze and its effect on the wind, temperature and humidity fields at coastal locations. The various

effects of orography and coastline curvature on the inland penetration of sea-breezes are also presented. More recent studies of the sea-breeze are, in general, the result of air quality research for coastal regions and the role of the sea-breeze in pollutant dispersal has been discussed.

Advances in computer power have been paralleled by an increasing sophistication in numerical models of the sea-breeze. These models have allowed researchers to obtain a greater understanding of the sea-breeze and are now just beginning to be used by regulatory authorities for air quality problems.

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