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The sea breeze and gravity-current frontogenesis

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SUMMARY

The development of a sea-breeze *circulation* into a sea-breeze *front* is often observed late in the afternoon. Measurements of horizontal temperature and humidity profiles in a number of sea-breeze circulations, made using an instrumented light aircraft, are presented. On each of the four days for which data are presented an initial weak horizontal temperature gradient, extending over a distance of 5 to 10 kilometres, developed later in the day into a sharp front only a few hundred metres across. The data are discussed in terms of simple gravity-current frontogenesis models applied to the development of a sharp sea-breeze front from an initial diffuse transition between the air over the land and sea. Examination of the equations for frontogenesis, and the results of some previous laboratory experiments on gravity-current frontogenesis, lead to the proposal that, given sufficient contrast between land and sea temperatures, the formation of a sea-breeze front depends on the balance between convergent horizontal winds which act to generate a front, and turbulent convective mixing over the land which tends to prevent its formation. The differences in driving force, wind convergence and turbulent intensity are discussed for each day on which data are available and are shown to be consistent with the theory proposed.

1. INTRODUCTION

The mesoscale meteorology of coastal areas is often controlled by the differential heating of the land and sea surfaces during the day. The resulting onshore-directed sea breeze transports cool, moist air inland, often modifying the local winds and causing cloud formation and, on occasion, precipitation. Pollutants released near the ground in coastal regions are often limited in their lateral and vertical extent by the sea breeze. Pollutants injected above the sea breeze may exhibit little mixing and dilution until fumigation occurs and the contaminants are incorporated into the surface layer. The prediction of mesometeorological flows and pollutant transport in coastal regions, therefore, requires an adequate understanding of the structure and mechanisms of sea-breeze generation and propagation.

The basic sea-breeze circulation is well understood. Cool, moist sea-air flows in over the land, warming and mixing with the land-air as it does so, with the result that the temperature and humidity contrasts between the sea-air and land-air decrease with distance inland. There is often a weak return flow aloft, and the whole circulation is restricted in depth by the height of the mixed layer (about 1 or 2 km) and extends inland from the coast 10–50 km.

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Under some circumstances (to be determined) there is a radical change in the sea-breeze flow. At the point of maximum penetration the transition between the sea-air and land-air is very sharp and takes the form of a sea-breeze front (Simpson *et al.* 1977). The sea-air which arrives at the front is cool and has high humidity, indicating that it has not mixed significantly with the land-air (nor been heated by solar radiation). In these cases inland penetration of the sea breeze is much greater than in the case where no front forms, and may be as much as several 100 km.

The mesometeorology is quite different in these two forms of the sea breeze, as is the fate of pollutants released in the coastal zones. This paper addresses the question of why (and how) a sharp sea-breeze front occurs on some occasions and not on others. When a front is formed the sea breeze takes the form of a gravity current (Simpson *et al.* 1977). In particular, the densimetric speed (U_{sb}) of a gravity current, or the sea breeze in the absence of a gradient wind, is approximately given by

$$U_{sb} = F(g'H_{sb})^{1/2} \tag{1}$$

where $g' = g\Delta\rho/\rho$ is the effective buoyancy force, g is gravity and $\Delta\rho/\rho$ is the non-dimensional density difference between the gravity current (or sea-air) and the surrounding fluid (or land-air) and H_{sb} is the depth of the sea-breeze flow. The Froude number, F, is a constant of order 1 in calm surroundings. The structure of gravity currents, including the existence of lobes and clefts at the leading edge, a thick bulbous head with Kelvin–Helmholtz billows at the rear, and a thinner tail region, have all been observed in sea-breeze fronts (Simpson *et al.* 1977; Simpson 1982, 1987).

Sharp contrasts in temperature, humidity, and wind speed and direction have been observed in the head of the gravity current (Koschmieder 1936; Simpson *et al.* 1977; Kraus *et al.* 1990). Often these contrasts are observed to sharpen in the late afternoon and evening and are associated with more rapid and further penetrations inland (Simpson *et al.* 1977). The existence of sharp contrasts between sea-air and land-air are indicative of poor exchange between the air masses with clear implications for contaminant transport and mixing. When there is a diffuse sea-breeze boundary, relatively good contaminant exchange might be expected between the sea-air and land-air. Under these conditions high concentration fumigation events are unlikely to occur.

The development of the frontal region has been the subject of an increasing volume of literature (e.g. Linden and Simpson 1986; Simpson and Linden 1989; Hacker *et al.* 1990; Kraus *et al.* 1990; Arritt 1991). Some of the key features of sea-breeze frontogenesis, however, were observed by Koschmieder (1936), who suggested that the synoptic winds must oppose the sea breeze before a front will form. Sea-breeze fronts have been observed, however, even when there is an onshore gradient wind (Pedgley 1958). Simpson *et al.* (1977) suggested that the front was the result of an increased temperature contrast between the sea-air and land-air in the afternoon. Mathematical and laboratory models of gravity-current frontogenesis by Simpson and Linden (1989), however, suggest that frontogenesis is associated with curvature in the mean horizontal density gradient and that frontal sharpening occurs whenever the wind direction points toward the region of lesser density gradient. Under these conditions there would be a convergence in the horizontal wind leading to tilting of the isopycnals in the vicinity of the front, and ultimately to the development of a sharp density contrast between the sea-air and land-air.

Opposing this tendency for frontogenesis is turbulence, which acts to smooth density gradients. In the case of the sea breeze the turbulence is generated by the convective motions driven by the unstable buoyancy flux from the ground. Linden and Simpson (1986) showed that when these turbulent motions are strong enough they cause vertical mixing, thereby reducing the baroclinic circulation associated with the horizontal density gradients. The interaction of these two opposing effects is the subject of this paper.

A recent numerical simulation (Arritt 1991) identified conditions under which tilting of the vertical stratification into the horizontal was also a significant factor in frontogenesis. High-resolution numerical simulation has the potential to examine the mechanism of sea-breeze frontogenesis in detail but, until recently, adequate field data have been lacking to support such efforts. Kraus *et al.* (1990) have generated high-quality field data which provide detailed budgets of the components of equations for frontogenesis in Australian sea breezes that can largely satisfy that need, and these data formed a basis of comparison for the work of Arritt (1991). The work of both Arritt (1991) and Kraus *et al.* (1990), however, focus on sea breezes that have already developed significant frontal character with sharp scalar property contrasts in the horizontal.

In this paper we examine the development of a sharp front from an initial condition of a diffuse transition from sea-air to land-air. In particular, we discuss the applicability of simple gravity-current frontogenesis models as described in Linden and Simpson (1986) and Simpson and Linden (1989) to sea-breeze frontogenesis. We use data collected during intensive field studies of the sea breeze over southern England during the period 1973–75. Simpson *et al.* (1977) used part of this data set to investigate the structure of some distinct sea-breeze fronts. We have examined the data not previously analysed, paying particular attention to the period leading up to the formation of the front. The data consist of aeroplane and ground-based monitoring of temperature, humidity and wind speeds and directions. No direct turbulence measurements were made. Data from four different days were analysed in detail for the current work. These days were selected on the basis of the availability of aeroplane flights before the onset of a sharp sea-breeze front.

2. FRONTOGENESIS

Frontogenesis is the strengthening of density and other property gradients in the fluid and can be derived formally from conservation equations for that property. Consider a scalar property ψ such as, for example, pollutant concentration, temperature, density or humidity, with ψ_x being the horizontal gradient of this property. The evolution of ψ_x in time t is given by

$$\frac{\mathrm{D}}{\mathrm{D}t}\psi_x = \frac{\partial u}{\partial x}\psi_x - \frac{\partial w}{\partial x}\psi_z - \frac{\partial}{\partial x}(k\nabla^2\psi). \tag{2}$$

Here u is the mean velocity in the onshore (x) direction and w is the mean velocity in the vertical (z). Equation (2) considers only frontal development in the onshore direction and assumes uniformity in the y-direction taken to be along the (straight) coast. The first term on the right-hand side represents the frontogenesis resulting from the convergence of the horizontal wind field. The second term represents the rotation of the vertical-property gradients as a result of a horizontal gradient in the vertical-velocity field. The final term represents the horizontal gradient of the turbulent flux with eddy viscosity, k, and will not be specified further here. Kraus *et al.* (1990) write a similar equation and formally define this term.

Kraus *et al.* (1990) and Arritt (1991) estimate the contribution of each of these terms in a developed sea-breeze front. These terms contribute to both frontogenesis and frontolysis (weakening of the front) locally near a developed front. In a developing front, the vertical-property variations tend to be small, i.e. temperature and humidity within the sea-breeze layer tend to be well-mixed vertically with a time-scale much less than an hour (McRae *et al.* 1981). In addition, the vertical velocity in a sea breeze without a strong front fluctuates and shows no strong trends. Thus mean horizontal shear in the vertical velocity is unlikely to be the major factor driving frontogenesis.

In a developing front, turbulence is expected to inhibit frontogenesis through the smoothing of temperature and humidity profiles. Turbulent heating may also account for the relative smoothness of the temperature profiles observed by Kraus *et al.* (1990) despite the development of sharp humidity fronts. It is likely that mesoscale frontogenesis in the sea breeze is controlled by the balance of velocity convergence and turbulence. This is consistent with the simple model of gravity-current frontal development developed by Simpson and Linden (1989) in the absence of turbulence. In that model, flow, both internal and external to the gravity current, was considered to be solely the result of density gradients, and frontogenesis was shown to occur if the horizontal density gradient in the advancing gravity current exceeded that in the ambient fluid. For the case of piecewise linear density gradients in both the gravity current and ambient fluid, Simpson and Linden (1989) showed that the time τ_f required for a front to develop was given by

$$\tau_f = c\{g(\rho_{1x} - \rho_{2x})/\rho\}^{1/2} \tag{3}$$

where c is a constant and the density gradients ρ_{1x} and ρ_{2x} correspond to the (assumed constant) density gradients in the gravity current and ambient fluid, respectively. For a continuous variation in density gradient, the rate of frontogenesis was shown (Kay 1992) to be related to the curvature in the density gradient. For this case, Eq. (3) can be written

$$\tau_f = c\left(g\frac{\rho_{xx}L_f}{\rho}\right)^{-1/2} \tag{4}$$

where L_f is the length over which the curvature in the density gradient ρ_{xx} occurs. On the basis of laboratory experiments on a gravity current without turbulence, Simpson and Linden (1989) found that c in Eq. (3) is approximately 4. Turbulent mixing would be expected to slow the rate of frontal development, and this retardation was observed experimentally by Linden and Simpson (1986) by bubbling air through a saline water gravity current. The development of sharp density contrasts between the advancing gravity current and the ambient fluid was inhibited by the turbulence generated by the bubbles, and was not observed until the air injection was turned off.

The application of these laboratory gravity-current models to the sea breeze requires consideration of the synoptic-scale gradient wind that is being perturbed by the coastal winds. Generally the gradient wind is nonzero and previous studies (e.g. Koschmieder 1936) have indicated that frontogenesis is more likely under conditions of an opposing gradient wind. The data of Kraus *et al.* (1990) were taken under conditions of a strong opposing gradient wind of 9–12 m s^{-1}. The equation governing frontogenesis, neglecting the variable effect of vertical-velocity shear and the net frontolytic effect of turbulence, can be written as

$$\frac{\mathrm{D}}{\mathrm{D}t}\psi_x = \left(-\frac{\partial u}{\partial x}\right)\psi_x. \tag{5}$$

This equation suggests that frontogenesis is a first-order exponentially increasing process (for negative velocity gradient or convergence) with time-scale τ_f given by

$$\tau_f \sim \left(\frac{\partial u}{\partial x}\right)^{-1}. \tag{6}$$

Dimensionally, Eq. (6) suggests that the time-scale for frontogenesis is given by the length L_f of the developing sea-breeze front divided by the velocity difference across the front

$$\tau_f \sim \frac{L_f}{U_{sb} - U_g} \tag{7}$$

where the densimetric sea-breeze velocity (U_{sb}) is given by Eq. (1) and the gradient-wind velocity (U_g) is the velocity unperturbed by the sea breeze.

Note that the velocity of the front should be of the order of the sum of the densimetric velocity and the gradient wind, while the time-scale for frontogenesis is related to the difference between these quantities. Laboratory and field data on gravity currents with both concurrent and countercurrent ambient flows (Simpson and Britter 1980) suggest that the front velocity is approximately given by

$$U_f \sim 0.87 U_{sb} + 0.59 U_g. \tag{8}$$

As indicated previously, however, turbulence tends to slow frontogenesis by mixing the sea-air and land-air. Since convective conditions drive the sea breeze, the turbulence is likely to scale with the convective velocity scale

$$w^* = \left(\frac{gQH}{\rho c_p \theta_0}\right)^{1/3} \tag{9}$$

where Q is the surface heat flux, H is the height of the mixed layer, c_p is the specific heat at constant pressure and θ_0, is the reference temperature of the air. The time τ_m required to mix the sea-breeze air over the entire mixed layer is of order H/w^* (Willis and Deardorff 1974). A dimensionless quantity that characterizes the tendency for frontogenesis in the sea breeze is then given by

$$N_f = \frac{\tau_m}{\tau_f} \sim \frac{H}{w^*}\frac{\partial u}{\partial x} \sim \frac{H}{w^*}\frac{(U_{sb} - U_g)}{L_f}. \tag{10}$$

Rapid frontogenesis would be expected for large N_f. For small values of N_f, turbulence would significantly deter frontogenesis and for $N_f < 0$, no frontogenesis would be possible. Because the relationship given by Eq. (10) is approximate, it is not possible to state *a priori* what specific values of N_f result in frontogenesis.

To examine the process of frontogenesis in the sea breeze, and to test the simple gravity-current view that the balance between turbulent mixing and convergence in the horizontal wind field largely controls frontogenesis, a series of field experiments conducted between 1973 and 1975 was re-analysed. Four experiments were selected on the basis of their potential for frontogenesis as characterized by the frontogenesis number, N_f.

3. Experimental methods

The data available for analysis were collected as part of an intensive field experimental programme conducted between 1973 and 1975 in southern England. Some of the observations have been discussed previously (notably by Simpson *et al.* (1977) and Simpson and Britter (1980)). The focus of the experimental programme was on the structure of the sea-breeze front and its penetration inland. The general study area and data collection points are shown in Fig. 1. Relative humidity, temperature and wind direction were monitored at ground stations at Horndean, Butser, East Tisted and Lasham. These ground stations were supplemented by an existing data collection network

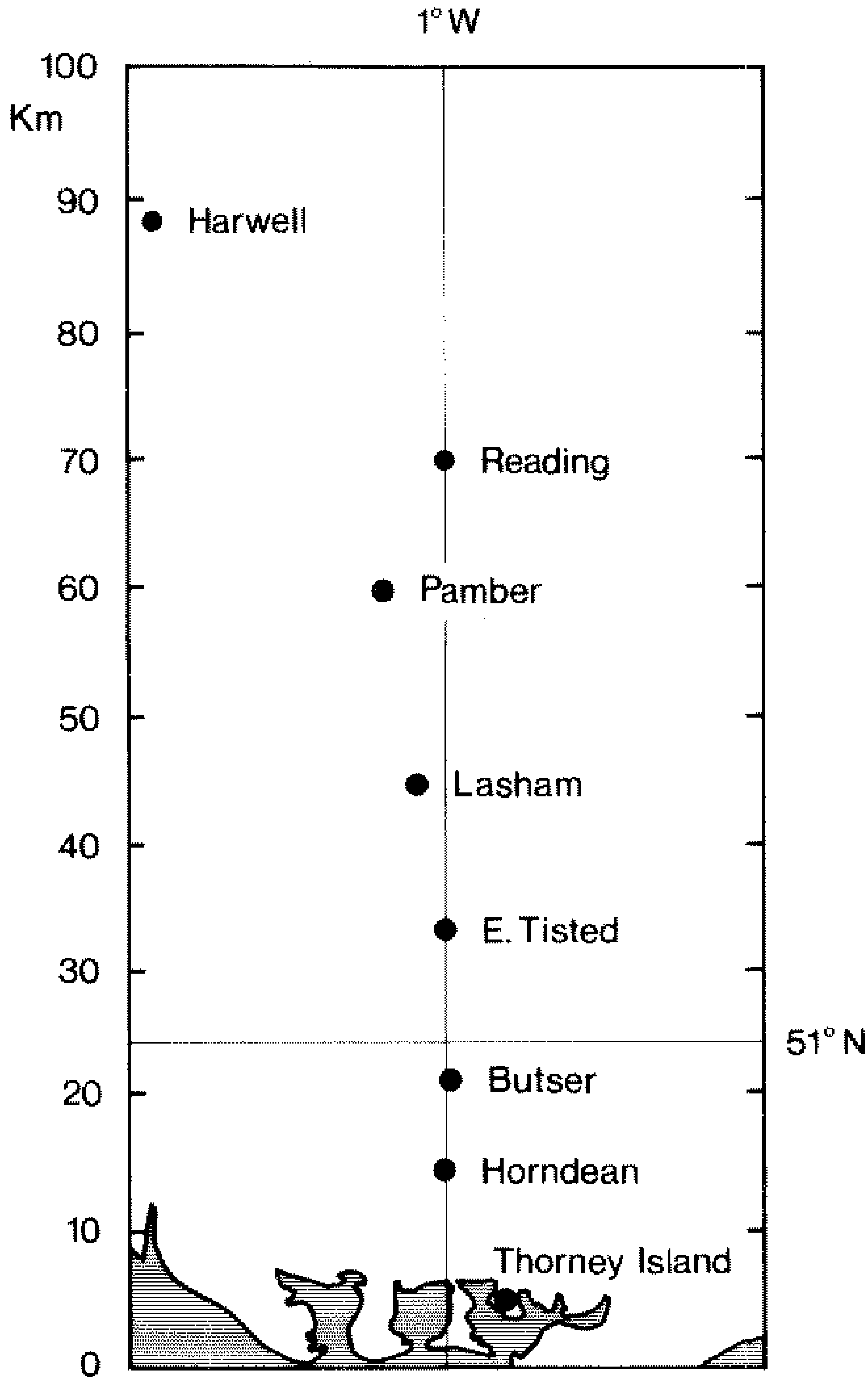

Figure 1. Study area in the south of England, showing the more important points for data collection.

that, in addition to these parameters, measured wind speed at Reading, Harwell and various other sites in the study area. Pollutant measurements, notably ozone, and solar radiation measurements were also available at Harwell.

Ground-based sites suffer, however, in that temperature measurements are usually influenced by a surface super-adiabatic layer that is not indicative of the entire sea-breeze layer. In addition, the near ground conditions are more heavily dominated by thermally and mechanically generated turbulence than conditions at altitude. Finally, much of the structure of interest at a sea-breeze front occurs at altitude. Thus the ground-based stations were supplemented by an instrumented Slingsby Falke 'powered glider' that allowed measurement of temperature and humidity as well as flight parameters (speed, altitude and rate of climb) every 1.6 seconds. The aircraft could also be used to estimate vertical motions as detailed in Mansfield *et al.* (1974). The basic instrumentation employed was as described by Milford and Whitfield (1970). The speed was held constant at approximately $30\,\mathrm{m\,s^{-1}}$ resulting in a horizontal separation between measurements of about 50 m. (The data shown later in Figs. 3–7 are averages over four successive measurements and so have a spatial resolution of approximately 200 m.) In addition to the instrumentation described in Milford and Whitfield (1970), a Lyman-alpha humidiometer (Electromagnetic Research Corporation) was used to measure water-vapour density. The response time of the humidiometer was significantly shorter than the 1.6-second sampling interval. The calibration of the humidiometer was checked against an Assmann psychrometer before and after each flight.

The aircraft traversed through the presumed sea-breeze front perpendicular to the coastline. Most of the data presented previously (e.g. Simpson *et al.* 1977) focussed on traverses and flights where distinct fronts were observed. On several days, however, the initial flights were conducted earlier in the day (mid-afternoon) and observations indicated a differential between sea-air and land-air conditions but no sharply contrasting front. The data from these flights form the basis for the current work. The aircraft was maintained at constant attitude during these flights to allow it to follow the vertical motions in and near the front. Despite this, the traverses were generally maintained at approximately constant altitude and were continued through the entire transition region between the cool, moist air and the hotter, dryer land-air. Occasionally flights or portions of a flight were devoted to a vertical sounding to indicate the temperature and humidity profiles with depth in the sea-air or in the unperturbed land-air.

4. Presentation and discussion of results

Data are presented for four days. During three of these days, two or more aeroplane sampling flights were conducted, allowing identification of the structure of the sea breeze both before and after development of a sharp front. Only ground-based data were available on the other day, but it was possible to infer the key characteristics of the frontal development from these data and by comparison to the other days. Figure 2 shows the inland progress of the sea breeze during these four days. The points joined by the dashed lines show the first detection of the 'sea-air' where no sharp front was observed. The solid line shows the advance of the sea-breeze after a distinct front had formed. Table 1 summarizes the temperature and humidity differences observed on these days, along with the approximate depth, and densimetric and onshore component of the gradient-wind velocities. In the last two columns the difference between the densimetric and gradient wind and N_f, as defined by Eq. (10), are included as indications of the frontogenetic potential. The estimates of N_f assume $L_f \cong 5$ km. Measurements of the solar radiation at midday on 14 June 1973 and 4 June 1974 suggest $w^* \simeq 1.6\,\mathrm{m\,s^{-1}}$ but no direct measurements were available for the other days. However, measurements were taken at similar times of the year and, since w^* depends on the solar heat flux to the $\frac{1}{3}$ power (Eq. (9)), it seems reasonable to use $w^* \simeq 1.6\,\mathrm{m\,s^{-1}}$ on the other days. The sum of the densimetric and gradient wind is included as an indication of the front velocity.

Based upon the difference between the densimetric and gradient wind, and assuming that the surface heat flux was approximately equal on all days (i.e. L_f similar), the first and third experimental days should exhibit approximately similar rates of frontogenesis while the second should be significantly slower and the fourth significantly more rapid. Since $\partial u/\partial x < 0$, frontal development was possible on each day.

The first test day, 14 June 1973, had no significant gradient wind. Estimates of the mixing and frontogenesis time-scales suggested that frontal destruction by mixing was occurring twice as rapidly as formation by wind-field convergence during mid-afternoon. The second day, 13 June 1975, exhibited a weaker sea-breeze driving force at mid-afternoon, but a moderate opposing gradient wind suggested that the overall potential for frontogenesis was similar to that of the first day. On the third test day, 26 June 1975,

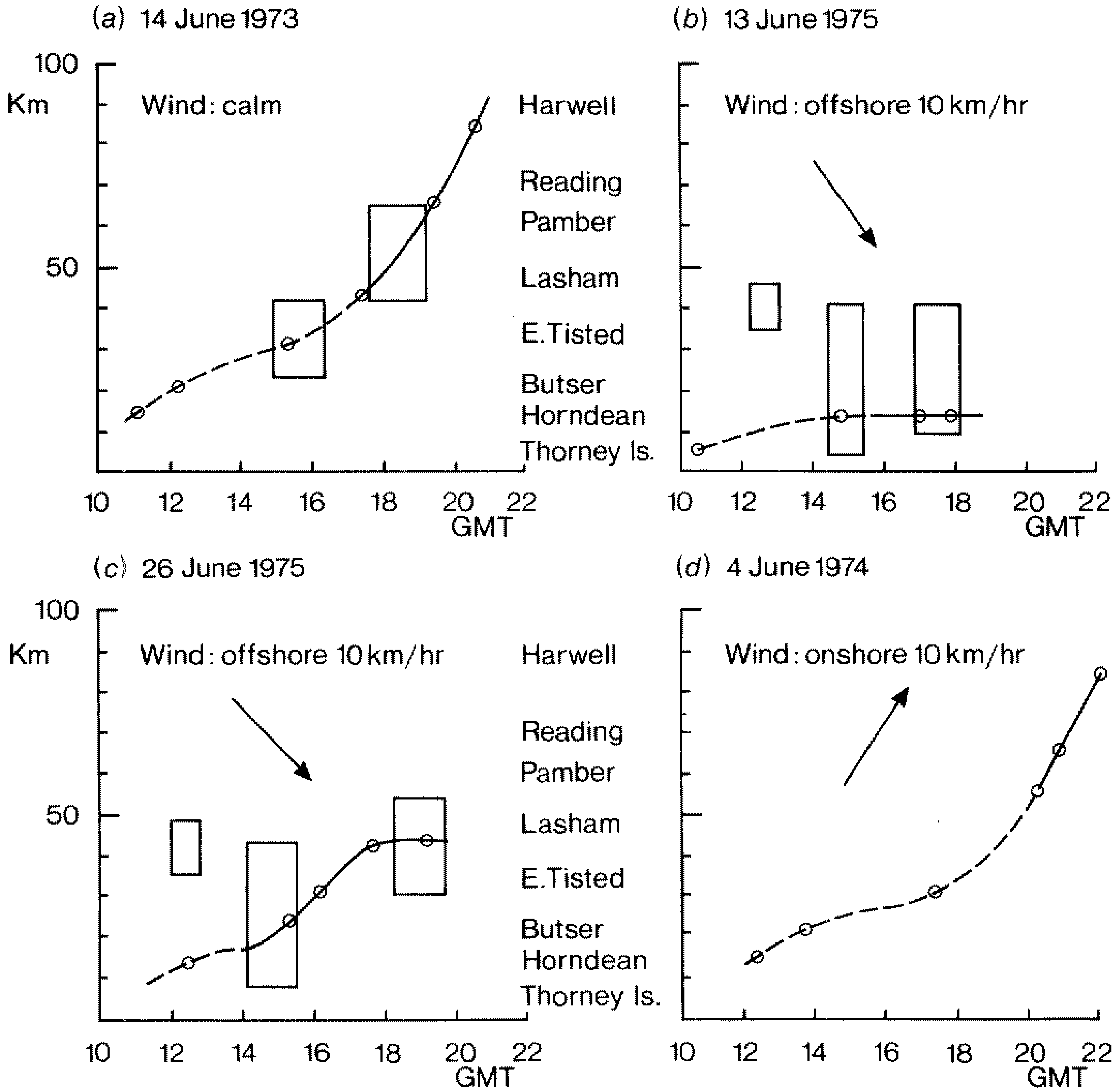

Figure 2. The generation and subsequent advance of the sea-breeze front on the four days investigated. The dashed lines show the first appearance of 'sea-air' and the solid lines the sea-breeze fronts. The boxes show the approximate times and places of the flights during which measurements were made. The arrows (drawn to scale) give an indication of the gradient wind on the four days.

the sea-air/land-air density difference was much greater and, in addition, an opposing gradient wind suggested that the conditions for frontogenesis were highly favourable. Finally, the fourth test day, 4 June 1974, was one of moderate sea-breeze potential as measured by the sea-air/land-air density difference, but a cocurrent gradient wind suggested that frontogenesis would be severely impeded. The sea-air extent as a function of time of day is shown in Fig. 2. Each of these days and the observed rate and character of frontogenesis will now be examined in detail.

(*a*) *14 June 1973*

Two airplane flights were conducted on this day, one between approximately 1500 and 1630 GMT and the other between 1730 and 1915 GMT. During the earlier flight, ground-based and airborne measurements suggested that the sea-air was essentially stalled in its progression inland. Sea-air had arrived at Butser (21 km inland) at about 1200 GMT but did not pass East Tisted, 13 km further inland, until about 4 hours later.

TABLE 1. CHARACTERISTICS OF EVALUATED SEA-BREEZE DAYS

Date	Δq g kg^{-1}	ΔT deg C	H_{sb} m	H m	U_{sb} m s^{-1}	U_g m s^{-1}	$U_{sb} - U_g$ m s^{-1}	N_f
14 June 1973	1	1	300–500	1200	4.2	<1	4.2	0.64
13 June 1975	1	0.5	200–300	750	2.6	−3	5.6	0.53
26 June 1975	2	2.5	300–500	1000	6.4	−3	9.4	1.25
4 June 1974	1	1	300–500	1100	4.2	3	1.2	0.18

The data shown were obtained during 1400–1500 GMT. The values of the mixing ratio, Δq, and potential temperature, ΔT, are representative of the overall differences between the land-air and sea-air. H_{sb} is the depth of the sea breeze and H is the depth of the mixed layer. U_g is the gradient wind estimated from surface observations inland, away from the influence of the sea breeze, and U_{sb} and N_f are defined by Eqs. (1) and (10), respectively. The values of Δq and ΔT for 4 June 1974 are depth-averaged humidity and temperature differences estimated by similarity of ground conditions.

Shortly after the end of the first flight, however, sea-air began to progress rapidly inland at about 15 km h^{-1} (4 m s^{-1}) as shown in Fig. 2. Simpson *et al.* (1977) attributed this acceleration to the development of a sharp front and the resulting intensification of the density difference at the front. The structure of the fully developed front is described in detail in Simpson *et al.* (1977).

Further examination of the earlier flight clearly shows frontal development between the two flights. Figure 3 shows humidity measurements collected during the two flights. Each curve represents the average of four traverses conducted at varying heights within 500 m of ground level. Each traverse was translated horizontally to align the transition between the sea-air and land-air. The reference location, indicated by 0 on the distance scale, was where the mixing ratio or humidity content of 5.2 g kg^{-1} was measured. The

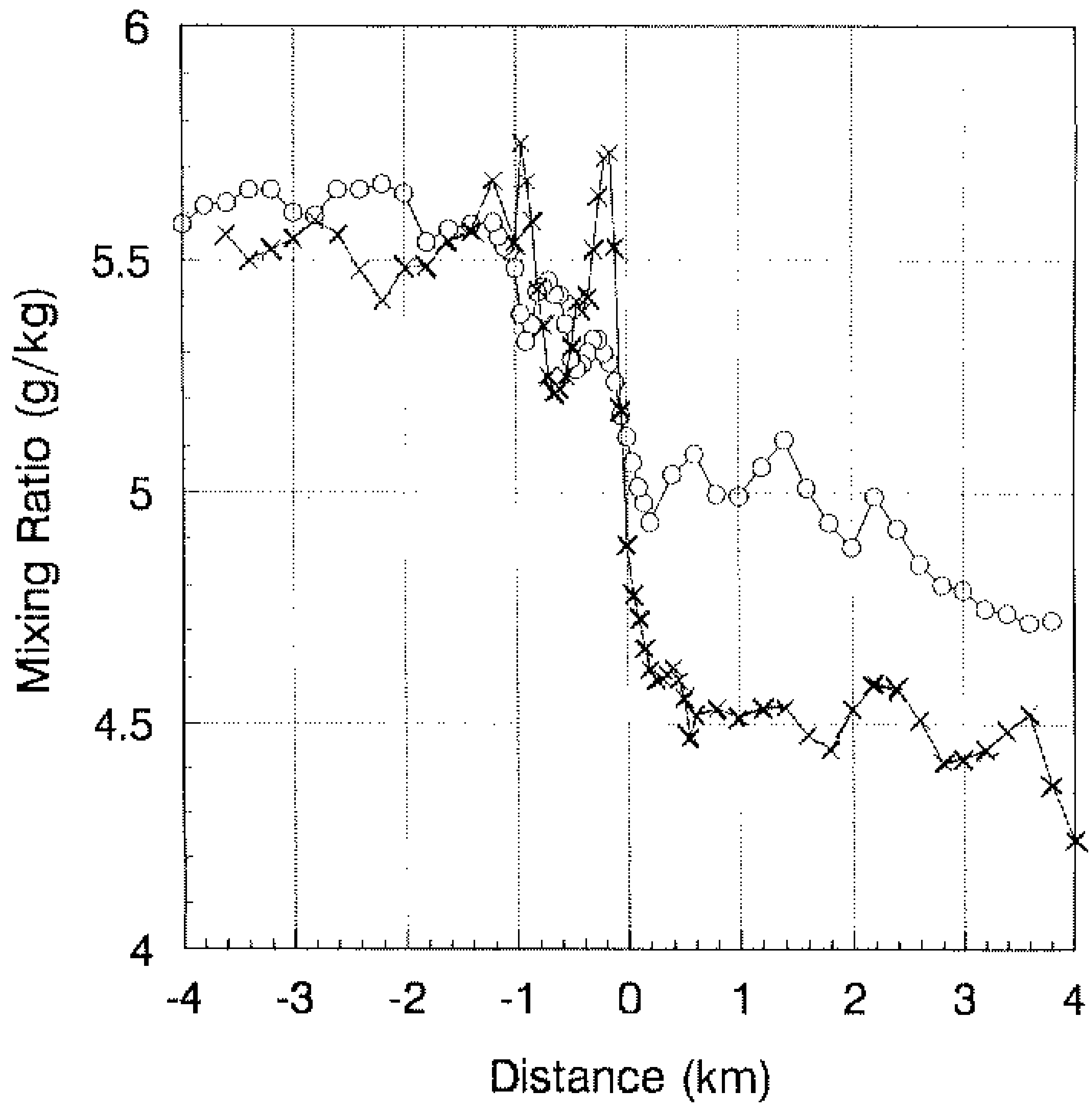

Figure 3. Humidity measurements during two flights on 14 June 1973. ○ 1505–1605 GMT, × 1730–1805 GMT. The traverses are moved to align the transition between sea-air and land-air.

earlier flight shows a transition between sea-air (containing about 5.6 g kg^{-1}) and land-air (containing less than 5 g kg^{-1}) that is spread over a horizontal distance of 4–6 km, consistent with our estimate of L_f. In the later flight, however, the entire transition requires less than 0.5 km. In addition to the clear development of a front, the data from the later flight show large-amplitude fluctuations consistent with the presence of Kelvin–Helmholtz billows immediately behind the front.

Figure 4 shows the clear differentiation between the sea-air and the land-air on the basis of humidity and temperature. On the first flight, the diffuse mixture of land-air and sea-air is clearly apparent from the continuous distribution of humidity and temperature at 1.6-second intervals. During the second flight, however, the transition is very distinct, and the humidity and temperature differences between the sea-air and the land-air far exceed the small fluctuations observed within either of the individual air masses.

Figure 5 shows that the temperature profile also sharpened considerably between the two flights. As with the humidity profiles, the existence of Kelvin–Helmholtz billows is suggested by the patch of warm air found immediately behind the advancing sea-breeze front during the second flight. Heating from the surface increased the temperatures in both the sea-air and land-air but the temperature difference between the air masses remained approximately 1 degC. The frontal development was apparently not caused by the intensification of the density difference between the sea-air and the land-air. Nor was frontogenesis the result of an opposing gradient wind, as the synoptic conditions were essentially calm throughout the day.

In the first flight, a temperature gradient in the sea-air of 0.28 deg C km^{-1} was observed. The temperature gradient in the land-air was an order of magnitude smaller. Although humidity is a good indicator of sea-air versus land-air, the density gradient is

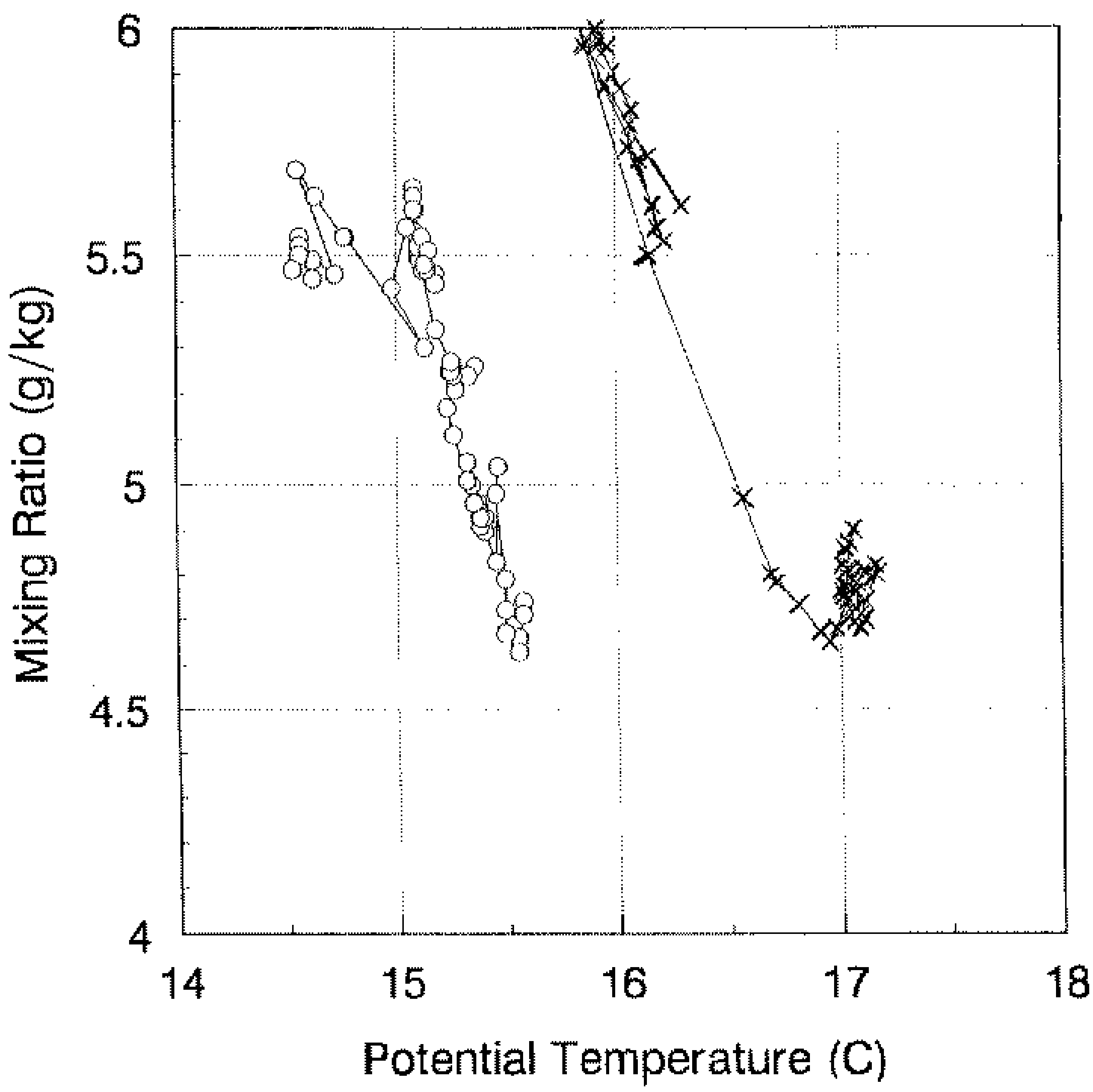

Figure 4. Measurements of humidity and temperature during the two flights on 14 June 1973. The second flight shows a distinct transition between sea-air and land-air. The symbols are the same as for Fig. 3. The traverses are moved to align the transition between sea-air and land-air.

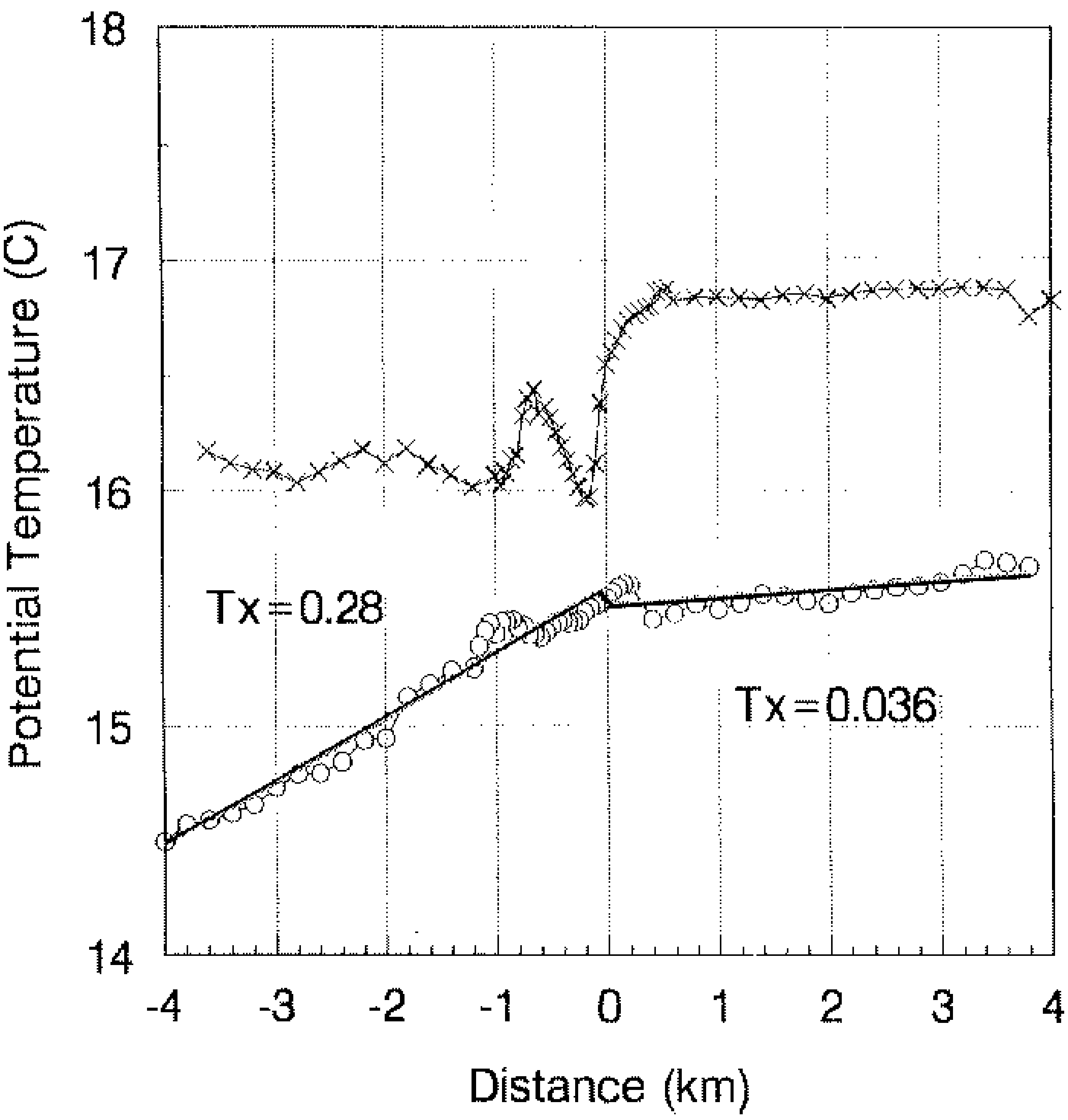

Figure 5. The sharpening of the horizontal temperature profiles between the two flights made on 14 June 1973. The symbols are the same as for Fig. 3. The solid lines are best-fit linear temperature profiles and the values of T_x represent the average horizontal temperature gradients on each side of the sea-breeze front. The presence of Kelvin–Helmholtz billows is suggested by the oscillations seen behind the advancing front in the second flight.

largely determined by the temperature gradient. The temperature gradient detected during the first flight is thus consistent with a decreasing horizontal density gradient, ρ_x, in the direction of the sea-breeze wind.

The model of Simpson and Linden (1989) suggests that since $(u\rho_x)_x < 0$, frontogenesis should occur with a time-scale given by Eq. (3) if the frontolytic effect of turbulence is neglected. If the small effect of the humidity changes on density is neglected, the temperature gradients shown in Fig. 5 suggest a time-scale for frontal development of about 21 minutes (with $c \approx 4$ in Eq. (3)). Over the 60 minutes required for the first flight, however, there was almost no frontogenesis. As indicated in Table 1 the ratio of the turbulent mixing time-scale to the time-scale for frontogenesis as defined by Eq. (10) was approximately 0.5 at midday. Thus convective turbulence should significantly inhibit frontogenesis. Estimates of incoming solar radiation at Harwell allow the sensible heat flux and w^* to be estimated by the Priestley–Taylor method (Stull 1988). At midday, w^* was approximately $1.6\,\mathrm{m\,s^{-1}}$ giving $\tau_\mathrm{m} \approx 0.2\,\mathrm{h}$. The estimated sensible-heat flux corresponded to a Bowen ratio of 0.2. The midday sensible-heat flux was estimated to be about 15% of the measured solar radiation of $800\,\mathrm{W\,m^{-2}}$. By 1800 GMT, however, with the incident radiation only 30% of its value at 1200 GMT, w^* was estimated to be about $0.8\,\mathrm{m\,s^{-1}}$ or less. This reduction in turbulent intensity was apparently sufficient to allow the observed frontogenesis to take place. It appears that N_f defined by Eq. (10) of the order of 1 would result in rapid frontogenesis. This rapid decay in convective mixing in the early evening hours is consistent with the simulations of Nieuwstadt and Brost (1986) who found that the turbulence decayed on a time-scale equal to τ_m after heating had stopped.

(b) 13 June 1975

As measured by the value of N_f, the potential for frontogenesis on this day was similar to 14 June 1973. The density difference between the sea-air and land-air was weaker, giving rise to a densimetric velocity of only 2.6 m s^{-1} compared with 4.2 m s^{-1} on 14 June 1973. An opposing gradient wind of about 3 m s^{-1} ($U_g = -3$ m s^{-1}), however, provided additional convergence to offset partially the weaker onshore flow. The opposing gradient wind should slow the rate of penetration of the sea-breeze air ($U_f \sim U_{sb} + U_g$) to near zero but enhance frontogenesis ($\tau_f \sim U_{sb} - U_g$).

Temperature profiles measured during three traverses through the sea breeze are shown in Fig. 6. The circles represent temperatures measured earlier in the afternoon (1420 GMT) while the remainder of the data were obtained later in the afternoon. Only the data in the vicinity of the front are shown in the later flights. The zero distance point in the figure corresponds to the mid point of the sea-air/land-air transition as measured by the humidity profiles, and the coastline is located at approximately -14 km. The earlier temperature measurements show that the sea-air was heated by convection from the surface to essentially the land-air temperature within 5–7 km of the coast. Since the sea-air humidity levels (not shown) extended to 10 km (over Horndean), this shows that temperature is not a good identifier of the air mass.

During the later flight, the transition from sea-air to land-air, as measured by both temperature and humidity, was also over Horndean (see Fig. 2). That is, there was esentially no movement of the sea-breeze front between the flights, and no translation of the data profiles to align the sea-air transition is required as in Figs. 3 and 4. Considerable sharpening of the front did occur between flights, however. Again, considering only the temperature gradient as indicative of the density gradient in the sea-

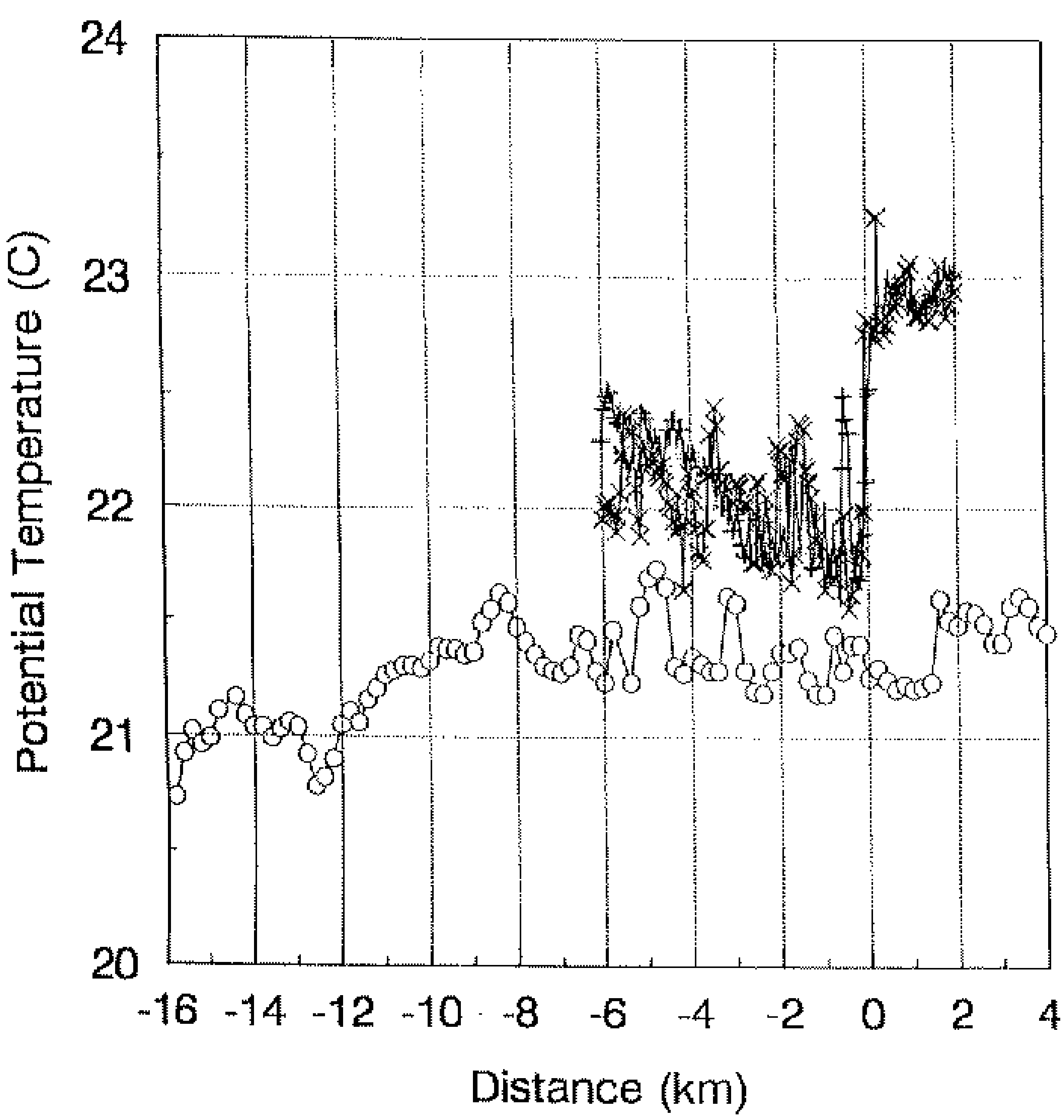

Figure 6. Temperature profiles made in three traverses on 13 June 1975. ○ 1420 GMT made at 270 m, × 1650 GMT made at 280 m, and + 1720 GMT made at 400 m height.

air, the maximum density gradient was only about half of the density gradient observed on 14 June 1973. The model of Simpson and Linden (1989) suggests that the time for frontogenesis would be 30% longer. This model, however, does not take into account the opposing gradient wind which increases the convergence of the horizontal wind field, encouraging frontogenesis. Since the opposing wind is approximately equal to the densimetric speed of the sea breeze, the convergence as a result of the gradient wind is approximately double that expected from the sea breeze alone. Thus the potential for frontogenesis should be approximately the same on 13 June 1975 as it was on 14 June 1973. That is, frontogenesis would again be expected when the midday turbulence levels decreased during the late afternoon and evening. As shown in Fig. 5, a sharp temperature front was first observed within 40 minutes of its observation on 14 June 1973. Without additional flights and direct measurements of w^* or solar radiation on this day, it is not possible to determine if this time difference is significant.

The late afternoon temperature profiles shown in Fig. 6 also show evidence of significant mixing behind the front. In addition to evidence of billows there are indications of a rapid increase in temperature to within 0.5 deg C of that expected for land-air within 6 km of the front. On this particular day, the midday mixing height was especially shallow (approximately 750 m at Lasham). Shear at the top of the gravity current tends to increase if an opposing flow is limited to a shallow layer (Simpson 1987).

(c) 26 June 1975

June 26 1975 differed significantly from the previous days analysed in that the temperature and humidity differentials between the sea-air and land-air were significantly larger on this day. As a result the densimetric speed of the sea breeze exceeded 6 m s^{-1}. In addition, a gradient wind of about 3 m s^{-1} was opposing the sea breeze. Since the densimetric speed exceeded the opposing gradient wind, the sea breeze would be expected to progress inland throughout the day, as shown in Fig. 2, unlike the sea breeze of 13 June 1975. The gradient wind would, however, increase convergence at the sea-breeze front enhancing the potential for frontogenesis. The value of N_f was already approximately 1 at midday, which is about the same as when frontogenesis was observed on 14 June 1973. As shown in Fig. 7, this resulted in the development of a sharp front even before the earliest flight at 1440 GMT. The potential temperature measured during the traverse increased from 21 to 23 °C within 400 m of passing between sea-air and land-air. A temperature gradient toward the coastline was also measured during this flight, suggesting that a density gradient remained to drive additional frontogenesis. During a subsequent flight at 1510 GMT, the transition from sea-air to land-air temperatures and humidity occurred between two adjacent measurements, a distance of 200 m. Recall that the data points shown in Fig. 7 are averages of four individual measurements (i.e. 200 m averages) and therefore do not reveal the true sharpness of the front. Each of the measured profiles shown in Fig. 7 are shifted to align the front location with distance zero as in Figs. 3 and 4.

The sharpness of the front even before the reduction of turbulent levels in the late evening is consistent with the strong convergence expected at the front, and with the value of $N_f \approx 1$ required for rapid frontogenesis.

(d) 4 June 1974

On 4 June 1974 flight data were not available. Based on ground temperature and humidity measurements, the strength of the sea breeze would be expected to be similar to that observed on 14 June 1973. In contrast with 14 June 1973, however, the gradient wind on this day was in the direction of the sea breeze. Thus the sea breeze would be

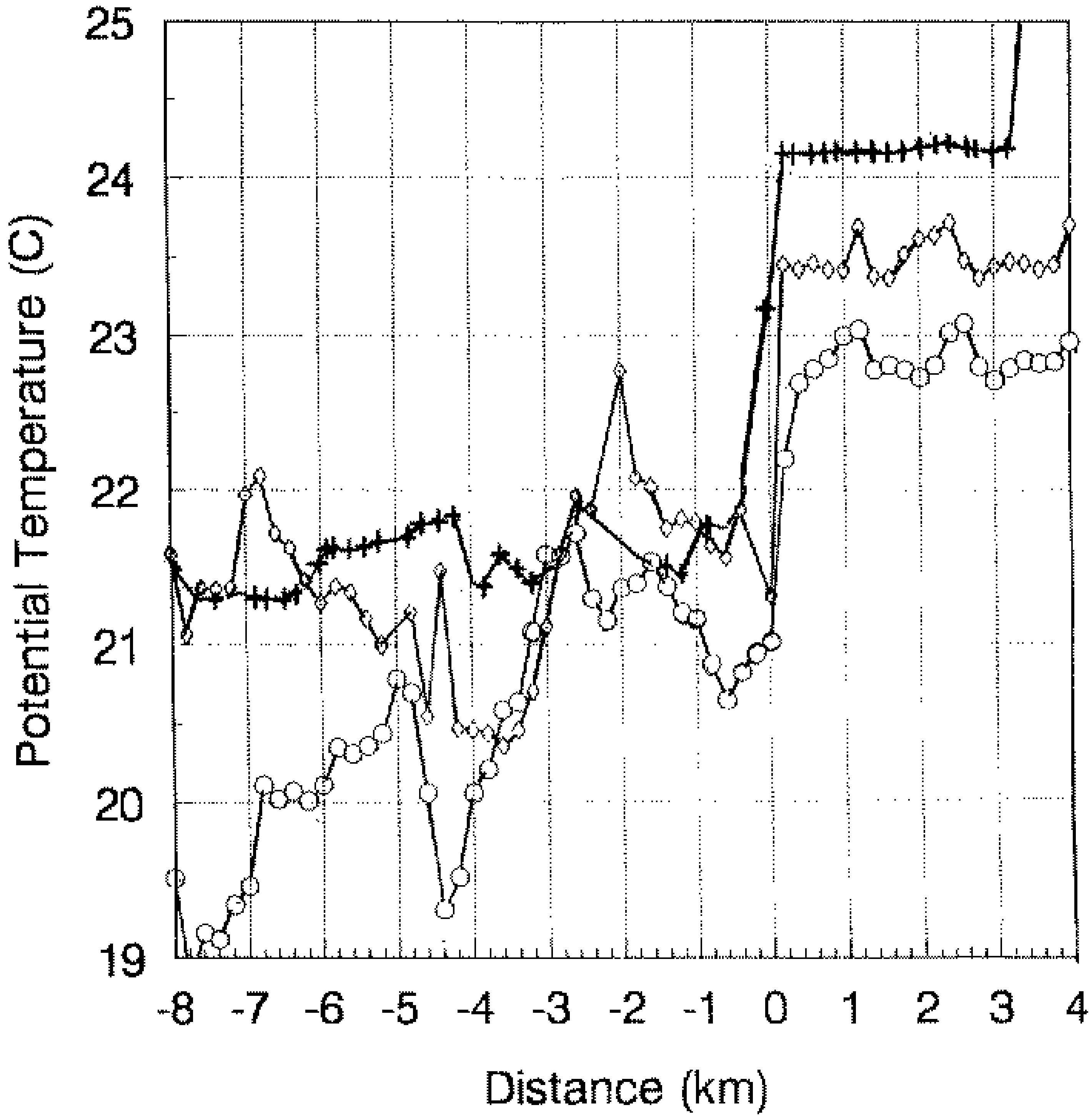

Figure 7. Temperature profiles made on 26 June 1975. ○ 1440 GMT made at 460 m, ◇ 1510 GMT made at 440 m, and + 1919 GMT made at 350 m height.

expected to penetrate rapidly inland at a rate proportional to the sum of the velocities, but that frontogenesis would be severely weakened by the lack of convergence in the wind field. Figure 2 shows, however, that the rate of rapid penetration of sea-air inland did not occur until about 2000 GMT much later, for example, than on the calm gradient wind day of 14 June 1973. There was no evidence of a sea-breeze front based on ground measurements of Lasham, 45 km from the coast, which is further inland than the point at which the sharp front had developed on any of the previous days. The lack of a sharp change in humidity or temperature combined with a southerly wind flow throughout the afternoon precluded any observation of a distinct transition to sea-air. The first indications of a front were at Pamber, near Reading, at about 2000 GMT, well after significant frontal development on the previous days. A distinct front was clearly detected at Harwell, about 90 km from the coast at 2140 GMT. By this time, convective conditions had largely ceased and turbulence in the surface layer significantly decreased. The sharpness of the front upon arrival at Harwell can be seen in Fig. 8. Onset of the sea-breeze front is associated with a rapid change in wind direction to southerly and simultaneous increase in ozone levels. Reduction in turbulence allowing the development of a strong sea-breeze front is indicated by the reduction in magnitude of the wind direction fluctuations before the detection of the front. Ozone would normally be destroyed by oxides of nitrogen and other scavengers in the evening and night-time hours, and this reduction in ozone concentration occurs at Harwell before the onset of the sea breeze. The rapid increase in ozone with the arrival of the sea-breeze air indicates that there may be fewer scavenger species in the sea-air and illustrates the lack of mixing at the sea-breeze front.

Although no quantitative prediction of frontogenesis is possible on 4 June 1974, the late frontal development is consistent with the lack of convergence in the wind field and

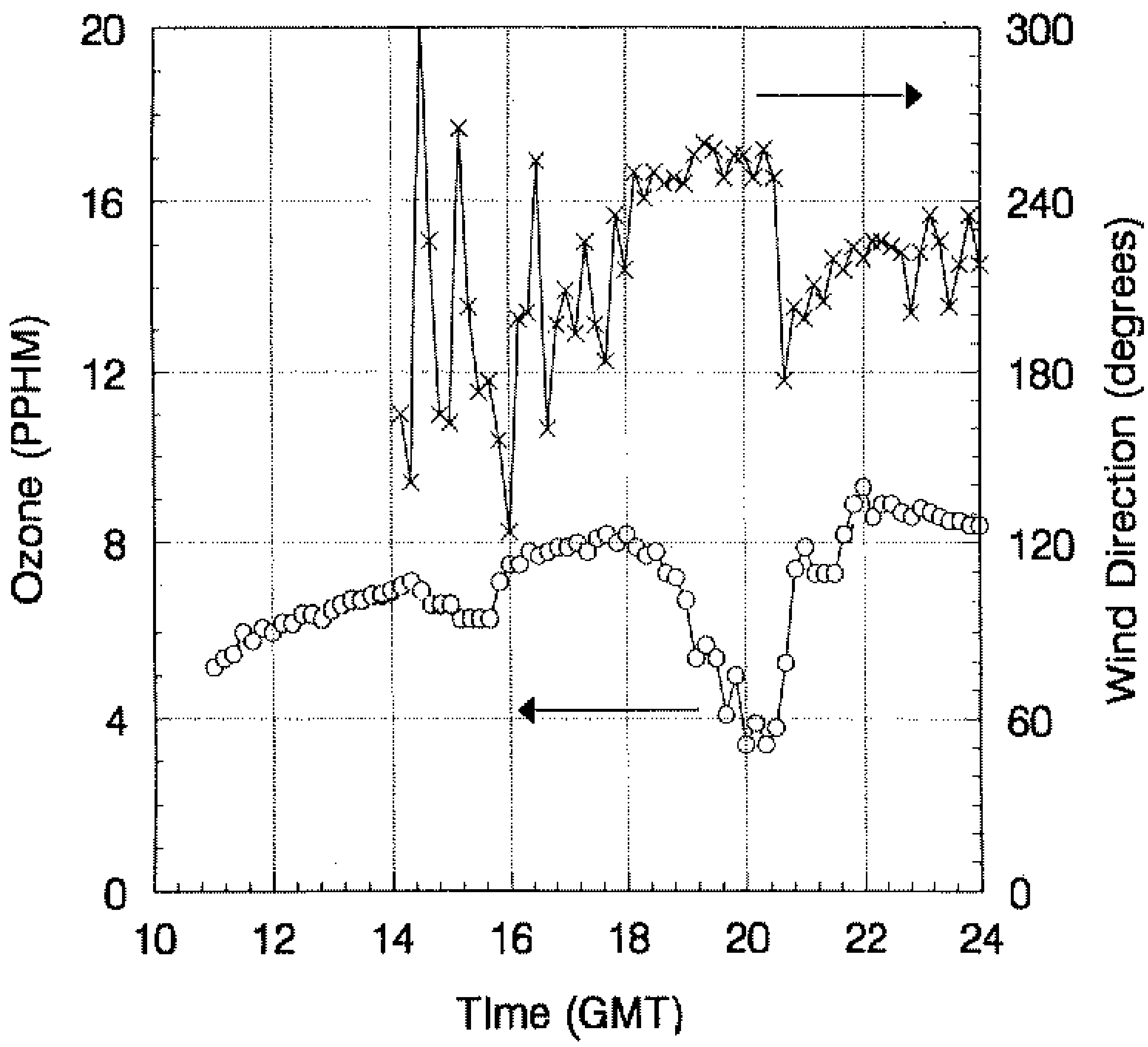

Figure 8. Measurements made at Harwell in the evening of 4 June 1974. The wind direction × is shown in degrees and the ozone concentration ○ in parts per hundred million (PPHM). The front is seen clearly at 2015 GMT, both by the wind direction shift and the rapid increase in ozone. The arrows point to the relevant scales.

the resulting small values of N_f. The fact that frontal development did ultimately occur, however, is consistent with the fact that $N_f > 0$, i.e. the densimetric sea-breeze wind exceeded the gradient wind.

5. Summary and conclusions

We have studied the development of the sea-breeze front by examining its formation over the south of England on 4 days. Our results show that the generation of a sharp front from the general sea-breeze circulation results from the competition of two opposing processes. Frontogenesis is produced by convergence in the horizontal wind field and this convergence may be produced either by density driven flows, or by synoptic winds or a combination of the two. Opposing this tendency to produce sharp horizontal density gradients are the turbulent motions, associated primarily with the convectively driven turbulence over the land.

Since the land–sea temperature difference is related to the solar radiation, in the absence of synoptic winds these two opposing processes have the same driving force, and a delicate balance ensues. In the early to middle part of the day, the turbulent motions are generally strong enough to inhibit frontogenesis. In this case the sea-breeze circulation is confined to a region about 10 km from the coast. As the sea-air travels over this distance it is heated and eventually achieves the same density as the land-air. Fronts are only observed to form later in the day when the convective motions diminish as the solar radiation decreases.

This picture is modified by the presence of a synoptic gradient wind. An offshore (opposing) wind increases the convergence and enhances the possibility of frontogenesis and it is observed that the front then forms earlier in the day. The presence of the opposing wind also inhibits the inland penetration of the sea-breeze front despite the strengthening of the density gradient at the sea-air/land-air transition. Conversely, an onshore (following) wind inhibits frontogenesis and increases the rate of inland propagation of the front. Comparison of observations on 14 June 1973 with those on 4 June 1974 indicate, however, that frontogenesis and not the gradient wind was the more significant factor influencing sea-air propagation inland. In this paper we have added the effects of the synoptic wind on frontogenesis in a simple linear fashion. It is by no means obvious that a nonlinear process like frontogenesis can be treated in this way despite the fact that the observations appear to be consistent with this simple assumption.

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