A Criterion for Thermal Stratification in a Wind-Mixed System

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ABSTRACT

The onset of thermal stratification in an isohaline, wind-mixed water body is shown, by a simple model and observations, to be determined by the parameter u_*^3/hB', where u_* is the friction velocity of the air just above the water surface, h the water depth and B' a buoyancy flux. Defined as $B' = g\alpha(\rho C_p)^{-1} \times [Q_0 - 2Q_1(ch)^{-1}]$, where g is gravitational acceleration, α the coefficient of thermal expansion, ρ the density of water, C_p the specific heat of water at constant pressure, Q_0 the net surface heat input, Q_1 the solar radiation that penetrates the water column and c the extinction coefficient for Q_1 in the water, this buoyancy flux is the net buoyancy input to the water, less an amount due to solar radiation penetrating the water column. The transition from the well-mixed to stratified regime occurs when u_*^3/hB' falls below a value of approximately 6700. This is supported by observations from a lagoon 3 m deep where the complete formation and breakdown cycle of thermal stratification occurs over several hours. A value of 1.8 is found for the ratio of the rate of increase in potential energy of the water column due to wind mixing, over τv_*, where τ is the surface wind stress and v_* the friction velocity in the water near the air-water interface. The value of this ratio was obtained from measurements made in the lagoon where the effects of water heating were considered, as well as wind mixing, on changing the potential energy. The development of the simple stratification criterion allows some predictions to be made of the influence of turbidity on the thermal structure of a water body.

1. Introduction

This paper considers thermal stratification in a water body subject to atmospheric heating and wind-induced vertical mixing. In particular, the effects of solar radiation penetrating the water column and causing heating at the surface and internally, are considered. A simple criterion is developed to indicate the transition from the well-mixed to stratified state of an isohaline water body, and predictions are supported by observations from the very shallow (2–3 m deep) Coorong lagoon.

Atmospheric heating of a water body tends to produce thermal stratification, whereas vertical mixing, induced by the wind or tide, tends to destroy the stratification. Numerous investigators have considered the existence of thermal stratification to depend on the relative energetics of the above two opposing processes. Kitaigorodsky (1960) used the idea to obtain the ocean mixed-layer depth from a balance of heat input at the water surface and wind-induced mixing. More commonly, the energetics balance has been applied to tidally mixed shallow seas, where a criterion for the transition from the well-mixed to stratified regime is considered. Simpson and Hunter (1974), Fearnhead (1975) and Simpson *et al.* (1977) have considered thermal stratification in coastal waters around the British Isles and Garrett *et al.* (1978) have considered stratification in the Bay of Fundy and Gulf of Maine waters.

The studies cited above support the hypothesis that the presence of thermal stratification is dependent on the relative energetics of atmospheric heating and vertical mixing. However, each study has considered heating to take place only at the water surface. This paper considers wind-induced mixing and atmospheric heating at the water surface and exponentially decaying with depth, resulting in a stratification criterion that is dependent on water turbidity.

2. Theory

a. Wind-induced mixing

Kinetic energy from the wind is transferred to the water column and produces vertical mixing, as discussed by Kraus and Turner (1967). Accompanying the mixing is an increase in potential energy of the water column, and as a means of quantifying the process, Turner (1969) suggested that the rate of change of potential energy of the water column (dV/dt), due to wind mixing, is equal to a constant fraction of the kinetic energy input from the wind to the water. This can be written as

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$$\frac{dV}{dt} = m\tau v_*, \tag{1}$$

where τ is the wind stress and v_* the friction velocity in the water defined by

$$\tau = \rho v_*^2, \tag{2}$$

and where ρ is the water density. The constant m is defined as the ratio of the rate of increase in potential energy of the water column to the wind energy τv_*. It has been assumed that the kinetic energy input from the wind to the water is a constant times τv_*.

b. *Heating effects*

Here we determine the potential energy changes of a water column arising from heating at the surface and with depth. Consider a homogeneous water column of depth h subject to atmospheric heating $Q(z,t)$. It can readily be shown that the rate of change of potential energy required to mix the heat input and maintain the well-mixed state is

$$\frac{dV}{dt} = \frac{g\alpha}{C_p}\int_0^h \frac{dQ}{dz} z dz - \tfrac{1}{2} gh \frac{\alpha}{C_p}[Q(z = h) - Q(z = 0)], \tag{3}$$

where g is gravitational acceleration, z the vertical coordinate measured positive upward with the bottom at $z = 0$, α the coefficient of thermal expansion and C_p the specific heat of water at constant pressure. In the analysis to follow, values of $\alpha = 0.0002(°\mathrm{C})^{-1}$ and $C_p = 4.07\ \mathrm{J\ g^{-1}\ (°C)^{-1}}$ are used, appropriate for temperatures of ~20°C and salinity 20‰.

The depth dependence of the heating function can be written as (see Holloway, 1979)

$$Q(z) = Q_1 e^{-c(h-z)} + Q_2 e^{-c'(h-z)}, \tag{4}$$

where Q_1 is the component of the shortwave radiation that penetrates the water column and is exponentially absorbed, as described by the extinction coefficient c, and Q_2 comprises the remaining terms of the surface heat budget which are absorbed in an infinitesimal layer at the water surface. A large value of c' ($\approx 40\ \mathrm{m^{-1}}$, which is not critical), assures the required very rapid attenuation of Q_2. Solar radiation incident at the water surface encompasses a wide wavelength band, ~300–3000 nm. However, radiation with wavelength $\geqslant 900$ nm is absorbed much more rapidly than the shorter wavelengths (see, e.g., Neumann and Pierson, 1966). The value of Q_1 is then the portion of the solar spectrum at the water surface, between 300 and 900 nm. If S_p is the solar radiation just penetrating the water surface, Holloway (1979) estimates that

$$Q_1 = 0.685 S_p. \tag{5}$$

A constant albedo of 7% is assumed in order to calculate S_p from measurements of incoming solar radiation.

Substitution of (4) into (3) yields, assuming $e^{-ch} = 0 = e^{-c'h}$ and $h - 1/c' = h$, the rate of change of potential energy required to maintain isothermal conditions:

$$\frac{dV}{dt} = \tfrac{1}{2}\rho h B'. \tag{6}$$

Here, B' is a modified buoyancy flux defined as

$$B' = \frac{g\alpha}{\rho C_p}\left(Q_0 - \frac{2Q_1}{ch}\right), \tag{7}$$

where $Q_0 = Q_1 + Q_2$ is the total heat flux into the water.

The value of B' is the total buoyancy input less some part of this input due to solar radiation penetrating the water column, indicating that the buoyancy input that must be mixed to maintain isothermal conditions is less than the total input. Furthermore, the less turbid the water the less buoyancy there is to be vertically mixed. Eq. (6) reduces to the expression obtained by Simpson and Hunter (1974) when all the heating is at the water surface, i.e., $c \to \infty$.

c. *The criterion*

Thermal stratification is considered to form when the rate of change of potential energy required to maintain isothermal conditions [Eq. (6)] exceeds the rate of change of potential energy due to wind mixing [Eq. (1)]. Hence, stratification is expected when $\tfrac{1}{2}\rho h B' > m\tau v_*$, or when

$$\frac{u_*^3}{hB'} < \left(\frac{\rho}{\rho_a}\right)^{3/2}(2m)^{-1}, \tag{8}$$

where ρ_a is the density of air and u_* the friction velocity of the air defined by

$$\tau = \rho_a u_*^2. \tag{9}$$

The problem remaining is to determine m and hence the right-hand side of (8).

It should be noted that the criterion is invalid for times of negative B' (cooling) when vertical mixing would be enhanced by convective motions unaccounted for in this theory.

3. Determination of *m*

A number of estimates of the wind energy used in changing the potential energy of a water column have been reported in the literature. Some results are summarized in Table 1 together with estimated corresponding values of m. These are found using

TABLE 1. Results from various investigations on the amount of wind energy used in increasing the potential energy of a water column. A value of m is estimated for each result.

Investigator	Method	Results	$m = dV/dt(\tau v_*)^{-1}$
Turner (1969)	Observations of the deepening of the ocean mixed layer during strong winds. Two results are given.	$\frac{dV}{dt}(\rho u)^{-3} = 2.6 \times 10^{-8}$ and 1.3×10^{-8}	13.3 6.7
Denman and Miyake (1973)	Value gave best fit in a numerical model of ocean mixed-layer deepening.	$\frac{dV}{dt}(\rho_a C_D u)^{-3} = 1.2 \times 10^{-3}$ ($C_D = 0.0013$)	0.96
Halpern (1974)	Observations of the deepening of the ocean mixed layer during a storm.	$\frac{dV}{dt}(\rho_a C_D u)^{-3} = 3.9 \times 10^{-3}$ ($C_D = 0.0013$)	3.1
Kullenberg (1976)	Observations of deepening stratification in coastal waters.	$\frac{dV}{dt}(\rho_a C_D u)^{-3} = 1.8 \times 10^{-3}$ ($C_D = 0.0011$)	1.6
Kato and Phillips (1969)	Laboratory tank with initially a stable linear density gradient. Surface stress is applied via a mechanical stirrer.	$\frac{dV}{dt}(\rho v_*)^{-3} = 1.25$	1.25
Wu (1973)	Laboratory tank containing two stable homogeneous layers. A surface wind stress causes mixing.	$\frac{dV}{dt}(\rho_a u_*)^{-2} u_s = 2.4 \times 10^{-3}$	0.0374

(2), (9) and the bulk aerodynamic equation for wind stress,

$$\tau = \rho_a C_D u^2, \tag{10}$$

where C_D is the drag coefficient and u the 10 m height wind speed. In Table 1, u_s is the wind drift velocity.

The values of m show considerable scatter with the most inconsistent result given by Wu (1973). Richman and Garrett (1977) discuss the result of Wu and suggest the inconsistency is due to the differences between the developing wave field in a laboratory tank and in the ocean. Variations between results are also discussed by Niiler and Kraus (1977), who suggest that in the ocean, mixing may be enhanced at times by internal waves or by inertial currents in the mixed layer.

A value of m is estimated from measurements made in the Coorong where in such shallow water (3 m) it was found essential to account for the effects of water heating, in addition to wind mixing, on changing the potential energy of the water column.

Consider a water column, depth h, with a density profile $\rho_0(z)$ and potential energy

$$V_0 = \int_0^h g\rho_0(z)zdz. \tag{11}$$

Over a small time internal Δt, the density profile $\rho_0(z)$, and hence the potential energy, will be changed by both heating $Q(z,t)$ and wind mixing $m\tau v_*$. Although the two processes act simultaneously, here it is considered that over the small time interval they act independently. First, the heating causes a density decrease $\Delta\rho(z)$ to form a profile of density $\rho_1(z) = \rho_0(z) - \Delta\rho(z)$ with potential energy V_1. Wind mixing then changes the density $\rho_1(z)$ to form a profile of density $\rho_2(z)$ with potential energy

$$V_2 = \int_0^h g\rho_2(z)zdz. \tag{12}$$

Therefore, over the time interval Δt, the potential energy change $V_2 - V_1$ is due to wind mixing, while the change $V_1 - V_0$ is a result of the heating. Considering the change due to wind mixing, it can then be shown, using (1) and (4), that

$$\frac{V_2 - V_0}{\Delta t} + \frac{g\alpha}{C_p}\left(hQ_0 - \frac{Q_1}{c}\right) = m\tau v_*. \tag{13}$$

The first term of (13) represents the observed rate of change of potential energy, while the second term is a correction factor due to water column heating. A value of m can be estimated using (13) by observing a deepening stratification and knowing the heating terms Q_0 and Q_1 and the values of h and c. The wind energy τv_* can be estimated from the wind speed using (2) and (10).

The formulation of the correction term in (13) is not exact. The solar spectrum incident at the water

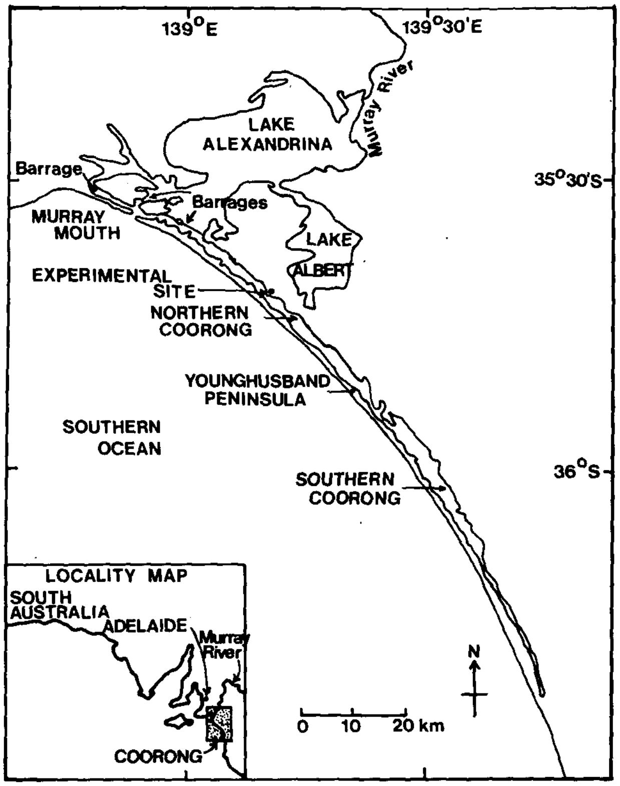

FIG. 1. Map showing the location and general features of the Coorong and the location of the experimental site.

surface is a continuous function of wavelength, whereas here it is assumed to consist of two distinct wavelength bands each with a uniform attenuation rate by water. Also, the separation of potential energy changes due to wind mixing and heating, rather than considering them as simultaneous processes, is an approximation. However, application of (13) to the shallow Coorong lagoon gives the correction factor to be of similar magnitude to the observed potential energy change term, making the inclusion of heating effects essential to the calculations.

4. Data

Measurements of varying temperature structure were made at one site in the Coorong, a 114 km long lagoon averaging 2–3 km in width and 2–3 m in depth, situated along the southeastern coast of South Australia (Fig. 1). The cross section of the lagoon at the measurement site is shown in Fig. 2. The southern end of the lagoon is closed, while the northern end is open to the sea and also receives a freshwater input from the Murray River, controlled by a series of barrages. With freshwater and saltwater input the measurement site, 30 km south of the northern end, experiences a wide range of salinity conditions. However, the times of stratification discussed here are for brackish water with vertically uniform salinity (~20‰). In addition, data sets were carefully chosen such that horizontal motions were wind and not tidally forced, and such that advection of heat was negligible. Heat balances of the cases considered show that all heating of the water column can be accounted for by vertical fluxes with horizontal transport neglected.

The water temperature profile was measured with a chain of ten thermistors, each with an accuracy of ±0.05°C. Wind speed was measured above the thermistor chain at 4.5 m height, using an Aanderaa cup anemometer, and corrected to a 10 m height. Net radiation, irradiance beneath the water surface, and current velocity and salinity at 0.6 and 2.0 m depths were also measured at the site. Incoming solar radiation data from Flinders University, ~95 km northwest of the site, was supplied by Forgan (personal communication). No correction for cloud was necessary as clear-sky conditions prevailed on the occasions selected here. Evaporative and sensible heat fluxes were calculated from bulk aerodynamic formulas and extinction coefficients calculated from the subsurface solar irradiance measurements (see Holloway, 1979). All variables were measured at 10 min intervals. Although the transfer coefficients used in the bulk

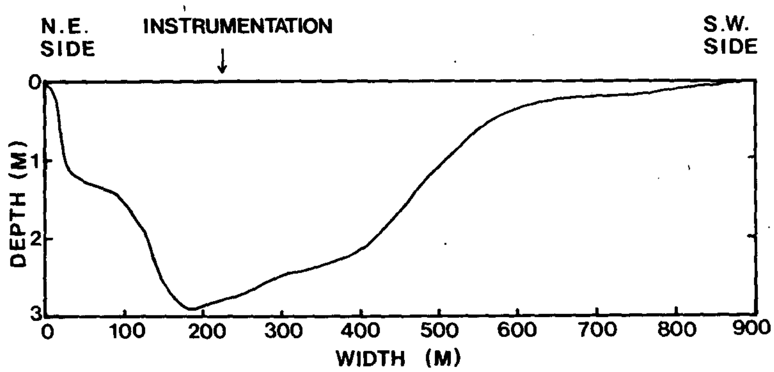

FIG. 2. The cross section of the Coorong at the experimental site and the position of the instrumentation.

aerodynamic formulas are those developed for large fetch situations, simple heat balance models for the measurement site indicate that the heat transfer processes, and hence the momentum transfer, are adequately described. This success may in part be due to the low wind speed conditions experienced in the analysis.

Three cases of thermal stratification are shown by the temperature profiles for 7 and 8 February and 4 March 1977, presented in Figs. 3a, 4a and 5a, respectively. The 10 min profiles are drawn from measurements at depths of 10, 50, 70, 90, 110, 140, 170, 200 and 240 cm, with an additional thermistor at 30 cm on 4 March. In each case stratification is seen to form from isothermal conditions under strong summer heating. The stratification strengthens, then, due to increased wind mixing (Figs. 3b, 4b and 5b), deepens and is finally destroyed, the complete cycle taking only a few hours. The stratification on 8 February (Fig. 4a) shows most clearly the formation and subsequent deepening of a surface mixed layer. Unfortunately, an hour's data is missing near the start of this series due to data logging tapes being changed.

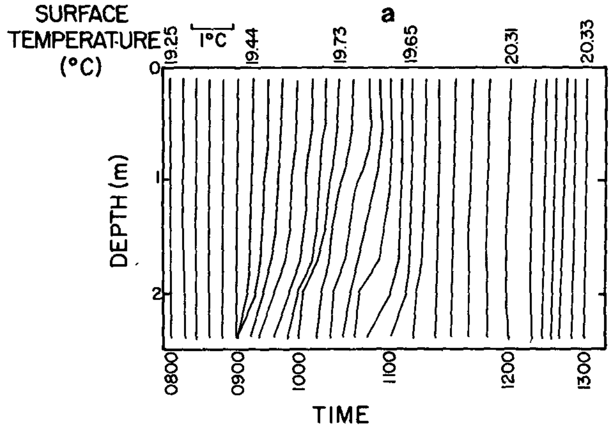

FIG. 3a. Water temperature profiles drawn at 10 min intervals from 0800 to 1300 local time on 7 February 1977.

5. Results

From the data described in Section 4, four values of m are calculated using (13), along with (2), (5) and (9)–(12). Table 2 summarizes the results, giving the time interval considered, h, c, and the average net heat flux into the water and average u_*^3 over the time interval indicated. The four values found (0.73, 1.2, 4.5 and 0.94) give an average of $m = 1.8$, although considering the range, we should not put too much emphasis on the average value. The time intervals were chosen such that significant change in the profile shape had occurred and also such that the observed total heat content change of the water column and the net surface heat input were approximately equal. It should be noted that, for the turbid waters, negligible radiation penetrates to the bottom of the shallow water column, validating the assumption that $e^{-ch} \approx 0$.

The major source of uncertainty in the calculated values of m comes from errors in the temperature measurements needed in calculating V_0 and V_2. With the temperature measurements accurate to ± 0.05°C, the resulting values of m have errors up to $\pm 45\%$, depending on the time interval chosen. This large error arises because the observed potential energy changes are small and hence sensitive to variations in temperature. However, this error is not sufficient to account for the range of values of m obtained.

The most inconsistent value is $m = 4.5$. Inspection of Figs. 3a, 3b and 3c shows that this large value occurs when the density interface that is being deepened is strong, whereas the lower values of m occur when a weaker density interface is being deepened. This observation supports, at least

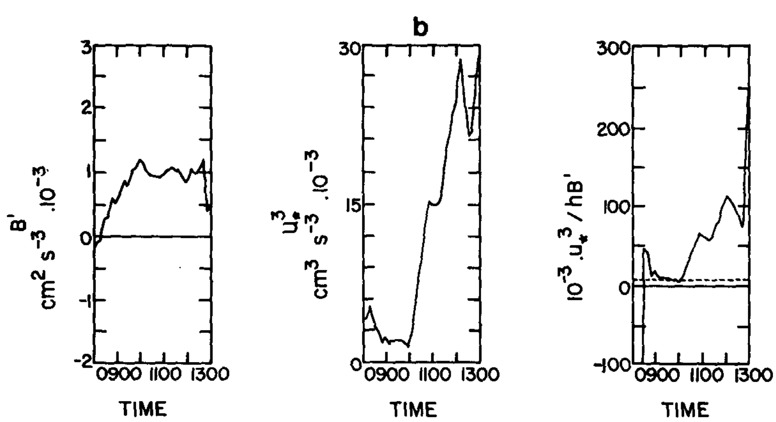

FIG. 3b. Time series of B', u_*^3 and u_*^3/hB' from 0800 to 1300 LT 7 February 1977. The dashed line shows the critical value of $u_*^3/hB' = 6700$.

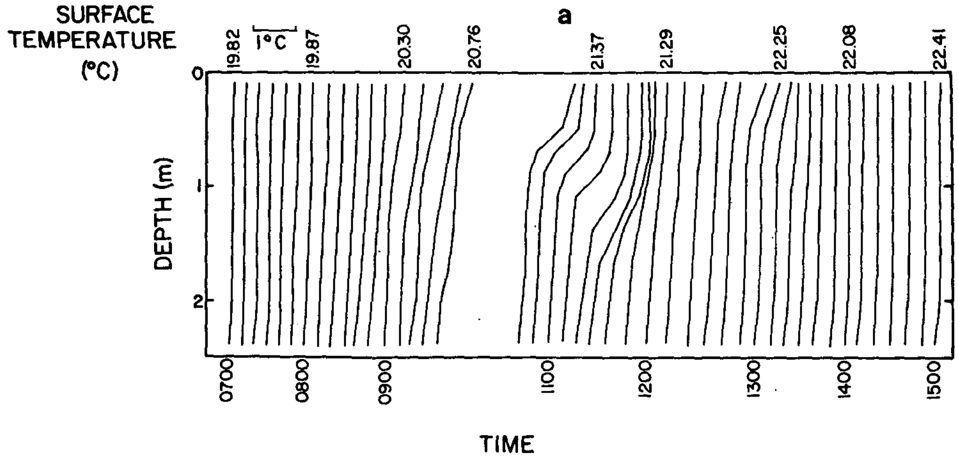

FIG. 4a. Water temperature profiles drawn at 10 min intervals from 0700 to 1500 LT 8 February 1977. The data from 0950 to 1030 are missing.

qualitatively, the hypothesis that m is not strictly a constant, but depends on the interface stability, as shown by the laboratory experiments of Kantha *et al.* (1977). Unfortunately, there are insufficient data to rigorously test this hypothesis.

The inclusion of the correction term due to heating in (13) was found to be essential as, typically, it was of similar magnitude to the observed rate of change of the potential energy term. Despite the range in the values of m (0.73–4.5), they do suggest that m is O(1), a result supported by other investigators (see Table 1). This agreement is especially pleasing considering the small length and time scales over which the observations of the deepening stratification were made.

Using the average value of $m = 1.8$, Eq. (8) predicts that a thermal stratification will exist when

$$u_*^3/hB' < 6700. \tag{14}$$

At this point it is convenient to define

$$S = u_*^3/hB' \tag{15}$$

as the stratification parameter, simplifying discussion in later sections.

6. Applying the stratification criterion to observations

The validity of (14) as a stratification criterion is tested by computing time series of S and comparing them against the observed stratification. This is done for the stratification shown in Figs. 3a, 4a and 5a, where values of h and c are given in Table 2, and also for an 8-day time series of measurements. The individual cases are discussed separately.

a. 7 February

Computed time series of B', u_*^3 and S are shown in Fig. 3b and should be compared to the temperature profiles in Fig. 3a. The dashed line in the time series of S represents the theoretically critical value of 6700. Thermal stratification is first evident at 0900 local time when S drops to near the

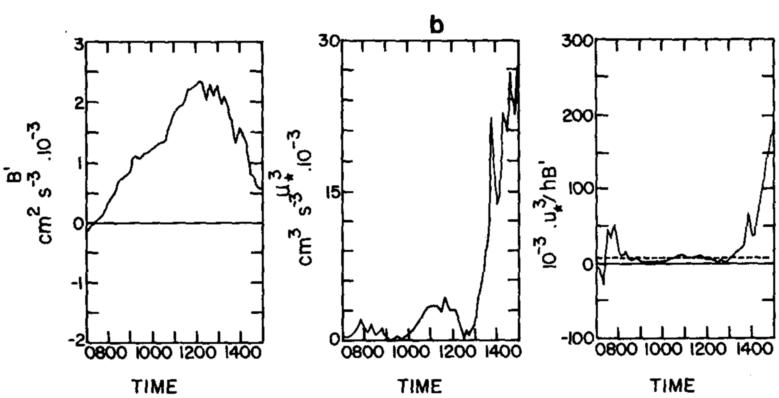

FIG. 4b. Time series of B', u_*^3 and u_*^3/hB' from 0700 to 1500 LT 8 February 1977. The dashed line shows the critical value of $u_*^3/hB' = 6700$.

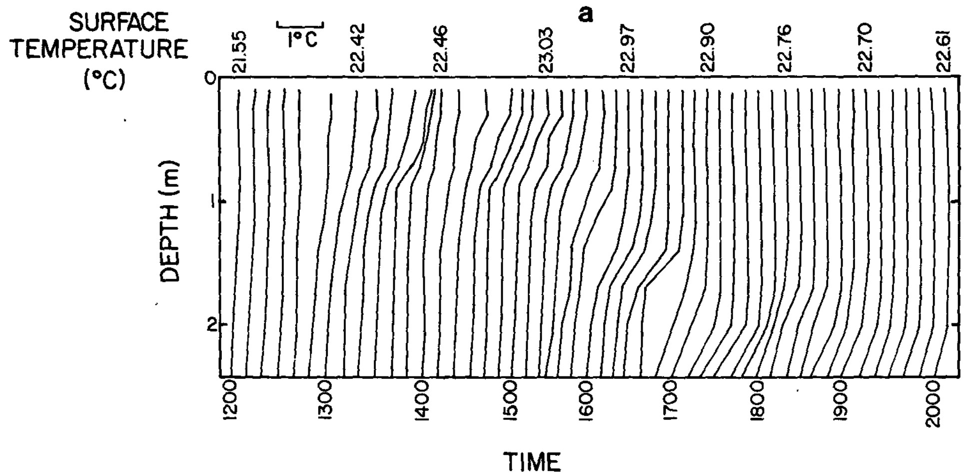

FIG. 5a. Water temperature profiles drawn at 10 min intervals from 1200 to 2000 LT 4 March 1977.

critical value due to an increase in B' and slight decrease in wind energy. It then intensifies as the value of S drops below 6700. At 1000 LT S rapidly increases, due to increased wind speed, exceeding the critical value. However, the stratification persists for another hour, being destroyed around 1110 when u_*^3, and hence S, increase further.

b. 8 February

Computed time series of B', u_*^3 and S are in Fig. 4b with the corresponding temperature profiles in Fig. 4a. Stratification commences forming at around 0800 LT, when B' becomes positive after the night cooling. This onset of stratification corresponds closely to $S = 6700$. The stratification strengthens for several hours, with S well below 6700, until around 1100 when the surface warm layer is significantly deepened corresponding to S increasing to around 6700. There is a dip in the value of S at around 1230, when a shallow warm layer is observed to start forming. However, the stratification is completely destroyed at 1310 when S rapidly increases, exceeding 6700. The variations in the depth of the stratification during this time series are well reflected by the variations in S.

c. 4 March

Computed time series of B', u_*^3 and S are in Fig. 5b with temperature profiles in Fig. 5a. No stratification was present in the morning due to the strong wind mixing (see Fig. 6). The onset of stratification at around 1250 LT corresponds to a drop in u_*^3 and is well predicted by S reducing to less than 6700. The stratification persists for several hours with S remaining below 6700 until 1530 when, due to decreasing B' and increasing u_*^3, S markedly increases. Although deepened, the stratification is not completely destroyed. A slightly cooler shallow bottom layer remains during the early evening when surface cooling has commenced. However,

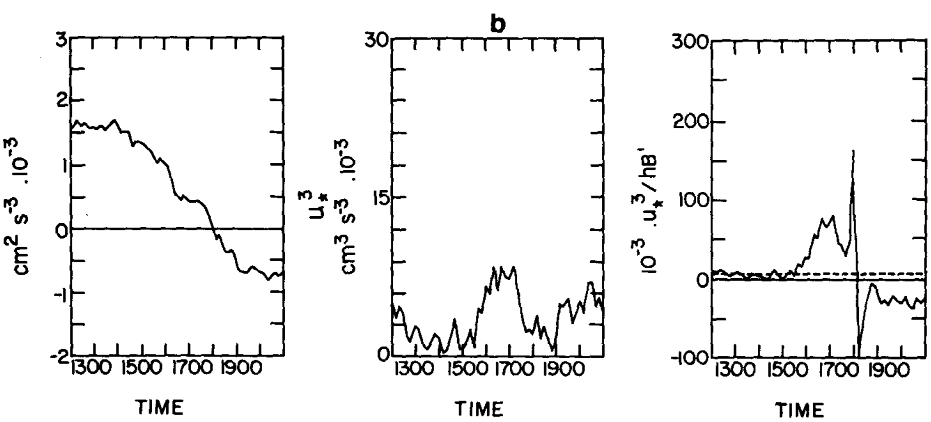

FIG. 5b. Time series of B', u_*^3 and u_*^3/hB' from 1200 to 2100 LT 4 March 1977. The dashed line shows the critical value of $u_*^3/hB' = 6700$.

TABLE 2. The values of m calculated from observations in the Coorong lagoon. Relevant parameters for each case are given where the overbar denotes an average over the time interval indicated.

Date	Time	h (m)	c (m^{-1})	$\bar{Q}_0$ (W m^{-2})	$\overline{u_*^3}$ (cm^3 s^{-3})	$m = dV/dt(\tau v_*)^{-1}$
7/2/77	1020–1100	2.5	1.7	396	12762	0.73
8/2/77	1040–1050	2.5	1.7	582	3096	1.2
8/2/77	1110–1200	2.5	1.7	762	3247	4.5
4/3/77	1430–1710	2.4	1.4	407	4792	0.94
					Average	1.8

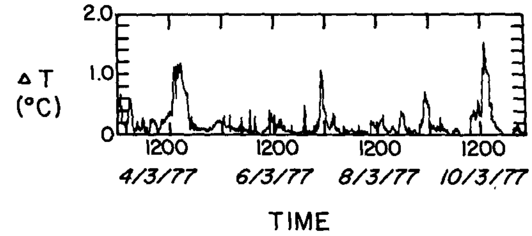

FIG. 7. Time series of the temperature difference between the 10 and 240 cm depths from 1200 LT 3 March to 0900 LT 11 March 1977.

this nighttime stratification may be artificial as the water depth was only 2.4 m, equaling the depth of the lowest thermistor. Consequently, this thermistor may have been slightly imbedded in the bottom sediment, thus giving a false reading.

d. 3–11 March 1977

The 8-day time series of S, from 1200 LT on the 3rd to 0900 on the 11th, is in Fig. 6 with the extent of stratification shown in Fig. 7 by a time series of the temperature difference between the 10 and 240 cm depths. Throughout this time, constant values of $h = 2.4$ m and $c = 1.4$ m^{-1} are used. The values of S are seen to vary over a wide range, from negative at night to values exceeding 3×10^5 during the day, two orders of magnitude greater than the theoretically critical value of 6700. Four major occurrences of stratification are seen on the 4th, 7th, 9th and 10th, each corresponding to $0 < S < 6700$, as predicted theoretically. At morning and night large oscillations are seen in the value of S as $B' \rightarrow 0$.

7. Discussion

The results discussed above clearly show a strong dependence of the presence and form of thermal stratification on the parameter S, defined by (15). Strictly, the development of the stratification criterion is valid only for the onset and not the breakdown of stratification and this is perhaps reflected in the results from 7 February and 4 March when the stratification persisted for longer than was predicted. This observation is supported by the laboratory experiments of Kantha *et al.* (1977), on turbulent entrainment across a stable density interface, where it is shown that as the jump across a density interface increases, more energy is required to produce mixing. Consequently, more mixing energy is required to destroy existing stratification than is required to prevent stratification from forming due to a heat input. In all the case studies from Section 6, the transition from the well-mixed to stratified state correspond closely to $S = 6700$.

The criterion for stratification is dependent on the value of m which for the calculated range (0.73–4.5) gives a variation in the critical value of S from 16500 to 2700. However, this is only a relatively small variation in comparison to the measured range of S (exceeding 3×10^5). Also, it can be seen that the critical value is close to the $S = 0$ line, giving a rather small range of S allowing stratifica-

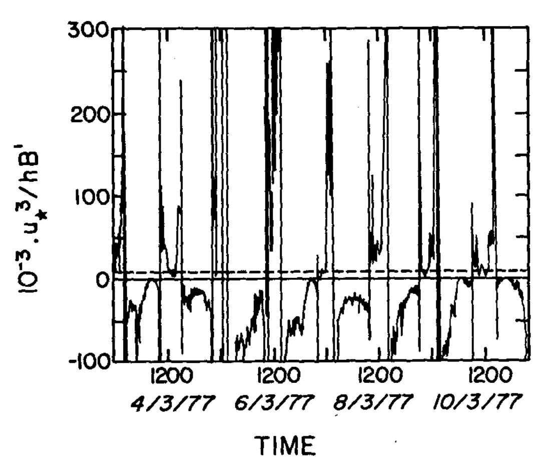

FIG. 6. Time series of u_*^3/hB' from 1200 LT 3 March to 0900 LT 11 March 1977. The dashed line shows the critical value of $u_*^3/hB' = 6700$.

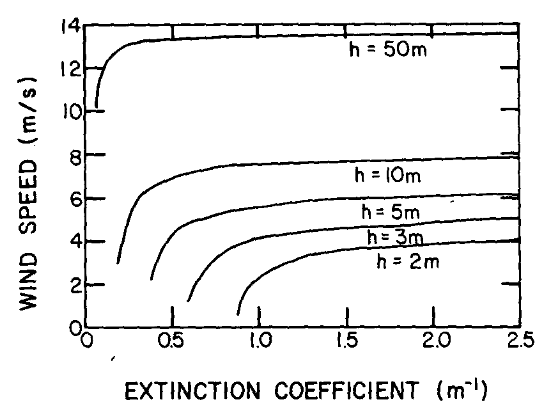

FIG. 8. Critical wind speeds below which a transition from the well-mixed to stratified state occurs, as a function of the extinction coefficient c, for varying water depths h. Values are obtained for typical summer conditions ($Q_0 = 700$ W m^{-2} and $S_p = 900$ W m^{-2}).

tion. Despite this, the results indicate that S drops to this range when stratification is forming. This is perhaps not a sensitive test of the numerical value of m, but it does support the use of the stratification criterion. It is difficult to assign an exact critical value of S, but the measurements support the theoretical value of 6700. It is concluded that (14) is a valid criterion for the onset of thermal stratification and that the parameter S reflects, to some degree, the depth of stratification.

It is of interest to make some predictions about the influence of turbidity of the water on the criterion for the onset of thermal stratification. Consider (South Australian) summer heating conditions of, say, $S_p = 900$ W m^{-2} and $Q_0 = 700$ W m^{-2}. The value of B' is then a function of h and c. By relating u_* to u using (9) and (10), Eq. (14) then predicts a stratification will form when

$$u^3 < 51h - 89/c, \qquad (16)$$

where MKS units are used. A family of curves for varying h are drawn from (16) to show the critical wind speeds below which a stratification will form, as a function of the extinction coefficient for the solar radiation. In Fig. 8 the curves show that as the water depth increases, or the turbidity increases, an increased wind speed is necessary to maintain isothermal conditions. For less turbid water the shortwave radiation penetrates the water column to a greater depth and so there is effectively less heat to be vertically mixed. Fig. 8 indicates that for each water depth there is a value of c below which the critical wind speed for stratification rapidly drops. Although this value of c decreases for increased h, it remains within the range observed in natural water bodies.

The results presented in this paper come from a rather simplified model of the "real" situation. The stratification parameter comes from a steady-state model in which convective stirring and many of the complexities of the partition of energy from the wind to the water have been neglected. Internal waves may have been present at the base of the surface mixed-layer and contributed to the mixing process, although it was not possible to tell from the measurements as a 10 min sampling interval was used. However, despite these simplifications to the model the success of the simple stratification parameter in predicting times of thermal stratification does indicate the dominant roles played by wind mixing and heat input through the water surface. Furthermore, the results indicate the influence of optical properties and of "internal" heating, due to solar radiation, on the stability and hence thermal structure of natural water bodies.

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