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Sea smoke and steam fog

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SUMMARY

The characteristics of fogs resulting from the advection of cold air over warm water (steam fog and sea smoke) are investigated. They are found to occur with air temperatures between 5 and 40°C lower than the water temperature, in winds from calm to gale; they have liquid water contents in the range 0·01-0·5 g m^{-3}, extend in height from 1 to 1,500 m and commonly exhibit either a columnar or banded structure. A study of their occurrence in Atlantic waters reveals a marked concentration in the western regions in the winter months because of the proximity of warm ocean and cold continent.

Through the use of equations for turbulent transfer it is shown that the occurrence of steaming is related to the well-known fact that two masses of unsaturated air at different temperatures when mixed together *can* yield a supersaturated or foggy mixture. In deriving the connexion the equality of transfer coefficients for heat and water vapour is assumed.

The circumstances in which steaming occurs are defined. The difference in the temperatures of the air and water must exceed a threshold which is dependent on the humidity of the air and the temperature and salinity of the water : its value, in the range 5-15°C, is a minimum when the air is moist and the water cold and fresh. The liquid water content and vertical extent of steaming increase as the thermal contrast increases. Close agreement is found between observations of the onset of steaming and the computed threshold values.

1. INTRODUCTION

The exchange of heat and vapour from a wet surface into the air overlying it is sometimes accompanied by 'steaming.' Wet, sun-warmed soil may steam; so may a body of water if the air above it is sufficiently cold. In the latter case the phenomenon is commonly termed steam fog (over fresh water) or sea smoke (over saline water)*. The major concern here is to define the circumstances in which steaming occurs : this is accomplished in Section 3 of the paper and shown to be correctly argued in Section 4. The theoretical discussion and examination of data is preceded by a summary of the distribution and physical characteristics of steam fogs since no current text is found to treat the topic in anything approaching a comprehensive manner.

2. DISTRIBUTION AND PHYSICAL PROPERTIES OF SEA SMOKE AND STEAM FOG

The steaming of natural waters takes place on all meteorological scales; micro, meso and synoptic. It occurs on a micro-scale when air cooled by nocturnal radiation drains from high ground onto a pond (Horton 1933; Woodcock and Stommel 1947). It occurs on a meso-scale when cold air cascades off the dark chill land mass of northern Norway onto the ice-free areas of the Fjord waters (Spinnangr 1949). It occurs on a synoptic scale when, in the wake of a depression, continental arctic air sweeps off the north-eastern coast of the U.S.A. out over the adjacent Atlantic (Brooks 1934).

(a) *Distribution of sea smoke (N. Hemisphere)*

Some features of the distribution of steaming on the synoptic scale were found by plotting about 60 reports of steaming in the Pacific and Atlantic Oceans. The following generalizations may be made : (*i*) Sea smoke commonly occurs outside polar latitudes and has been reported as far south as the tropics (Bannister 1948; Starbuck 1953) : thus the term arctic sea smoke seems particularly inappropriate; (*ii*) Over 90 per cent of the

* Other terms for the phenomenon are listed in the Appendix.

observations of sea smoke were made in the winter months, December to March; (*iii*) Because of the presence of warm water close to the western oceanic boundaries and because of the general eastward motion of cold continental air, the western parts of the ocean experience more frequent and more intense steaming than the corresponding eastern parts; (*iv*) Steaming is generally confined to coastal waters in low latitudes but with increasing latitude is reported at locations increasingly remote from the coast : Hay (1953) describes sea smoke at latitude 60°N after the air had an overwater trajectory of nearly 1,000 miles.

(*b*) *Factors influencing their occurrence*

Because it is well known that only when overlying air is *sufficiently* cold can a water surface steam, many investigators have looked for a threshold value for the difference in temperature between air and water. Thus Jacobs (1954) asserts that steaming never occurs in the Gulf of St. Lawrence when the air is less than 9°C colder than the water. Yet Horton (1933) observed steaming of a pond with a difference of only 6°C, and the author (see Table 1) has observed differences near 14°C in the absence of sea smoke. As demonstrated later, the relative humidity of the cold air is an important factor in this threshold value.

Observations of steaming have been made in winds ranging from near calm (Bryson 1955) to nearly 30 m sec^{-1} (Rodewald 1937; Spinnangr 1949), but according to Church (1945) its speed has little effect on whether a water surface steams or not.

(*c*) *Their vertical extent*

Steam fog and sea smoke are widely regarded as shallow phenomena : they are not. In deep cold air the height of sea smoke can exceed 1,500 m (Jacobs 1954; Berry, Bollay and Beers 1945; Cunningham (private communication)), and ship reports of steaming in excess of 100 m are not uncommon. The depth of steaming may on occasions be limited by the vertical depth or stratification of the cold air.

(*d*) *Their form*

The characteristic form of steam fog and sea smoke varies with the wind. In near calm Bryson (1955) observed an array of quasi-steady convergent columns with a spacing of a few meters and height of 2 m. Horton (1933) describes similar columns which were rotating; these had a diameter of $\frac{1}{2}$ m and height 5 m and drifted unsteadily across a river. On a larger scale, Brooks (1934) Woodcock (private communication) and others have described unsteady rotating columns of fog with vertical dimensions of 100 m or more, the scene likened to Dante's *Inferno*; Woodcock reported a wind of 5 m sec^{-1}. Church (1945) has used the terms sheet and blanket, indicating a well-marked top to the fog layer. The author's observations indicate that in moderate winds sea smoke commonly has a banded structure (see Fig. 1, Plate V) with the bands approximately along the wind, and it is a surprise to find only one other mention of this form in the literature (*Marine Observer* 1931, **8**, p. 60). In deep cold air steaming may be accompanied by cumulonimbus (*Marine Observer* 1957, **27**, p. 187; 1960, **30**, p. 12), cumulus or low stratocumulus.

(*e*) *Visibility and water content*

In sea smoke visibility as low as 30 m (*Mar. Obs.* 1959, **29**, p. 11), 50 m (Brooks 1934) and 100 m (Rubin 1958) has been reported : despite the widespread use of radar such obscuration represents a navigation hazard. On the other hand, steaming may be so slight and shallow that the visibility is not appreciably affected : in this case it is common to observe pronounced refractive shimmering and cusping of the horizon.

From the observations of visibility we can estimate the water content in steam fog as up to several tenths of a gram per cubic metre (Houghton and Radford 1938). The estimate is confirmed by direct observation of water content in winter fogs over the river Angara (in an industrial area) by Bashkirova and Krasikov (1958); these authors made a rough classification of fogs into tenuous, moderate, and dense, and found water contents of 0·03-0·04 g m^{-3}, 0·05-0·11 g m^{-3} and 0·08-0·37 g m^{-3} respectively. The water content was found to increase with increasing air-water temperature difference. In the cleaner air of the Arctic, in Kola Bay near Murmansk, water contents of 0·02-0·04 g m^{-3} and 0·04-0·14 g m^{-3} were reported in moderate and dense fogs.

(*f*) *Microphysical structure*

(1) Phase state

At air temperatures below 0°C the condensed water in sea smoke and steam fogs is commonly supercooled : riming of exposed surfaces in dense cold steam fog may thus be considerable (see Lee 1958; Mitchell 1958; and foreground, Fig. 1, Plate V). The Russian investigators of the steaming Angara River found that at temperatures above − 9°C to − 10°C the fogs consist of supercooled drops alone, whilst at temperatures below about − 20°C the condensate was predominantly ice. At intermediate temperatures fogs were mixed, with many spherical frozen drops. On occasions of steaming of Kola Bay, fogs remained supercooled to much lower temperatures, − 18°C to − 22°C, before crystals and irregular solid particles formed. Bashkirova and Krasikov interpret the differences as due to the industrial contamination of the Angara fogs.

(2) Drop sizes

One important feature of both the Angara River and Kola Bay fogs was that a decrease in air temperature was accompanied by a *decrease* in the size of the most frequent drop and a *decrease* in the width of the spectrum but an *increase* in water content; a similar behaviour was exhibited by the solid phase. Thus a reduction of air temperature resulted in the activation of not only greater numbers of freezing nuclei but also greater numbers of condensation nuclei. The writer believes that the latter result is a reflection of the increase in the rate at which air is brought to the condition of saturation with increasing air-water temperature difference.

3. The joint transfer of heat and water vapour

(*a*) *A necessary condition for the onset of steaming*

The simplest circumstances of steaming arise when *deep homogeneous* cold air overruns *fresh* warm water of *uniform* surface temperature. The potential temperature of air crossing the shore is denoted as θ_0, its mixing ratio as r_0. There is no restriction of the shape of the shoreline nor the variation of wind with height or time except in order to preserve unambiguity of the phrase ' downwind of the shore.' The equation for the *mean* value of a transferable property x is then written with the usual notation :

$$\frac{dx}{dt} = \frac{\partial}{\partial x}\left(K_x \frac{\partial x}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial x}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial x}{\partial z}\right) \quad . \quad . \quad (1)$$

where x is θ and r in turn, and K_x, K_y, K_z are the three components of the turbulent transfer coefficient. At the water surface ($z = 0$) it is supposed that the air takes up the

Figure 1. Sea smoke in Great Harbour, Woods Hole, Mass., U.S.A. 0915 EST, 31 December 1962. Water temperature − 0·2°C; air temperature (10 m) − 15°C, humidity 60 per cent. Wind NW 10 m sec^{-1}. Height of steaming approximately 5 m.

temperature T_F of the surface and the saturation mixing ratio for this temperature $r_v(T_F)$. Hence boundary conditions are

$$\theta = \theta_0, \qquad r = r_0 \qquad \text{at shore and} \qquad z > 0$$
$$\theta = \theta_F = T_F, \; r = r_v(T_F) \; \text{downwind of shore and} \; z = 0 \qquad . \qquad (2)$$

If the transfer coefficients for heat and water vapour are everywhere equal then it follows from Eqs. (1) and (2) that θ and r are everywhere linearly related; that is

$$r - r_0 = \beta(\theta - \theta_0) \quad . \quad . \quad . \quad . \quad . \quad . \quad (3)$$

where

$$\beta = \frac{r_v(T_F) - r_0}{\theta_F - \theta_0} \text{ is a constant} \quad . \quad . \quad . \quad . \quad . \quad (4)$$

This simple and powerful result holds however complex is the dependence of wind and transfer coefficients on space and time; it fails, however, if the transfer coefficients for heat and water vapour are unequal. Although differences have been reported (see Priestley 1959) they are believed to vanish close to the surface where the Richardson number is small. Since the major fraction of the temperature-drop in cold air over warm water occurs (in the mean) below a height of about 10 cm, the assumption of equality should be good. The roles of molecular conduction and diffusion, whose coefficients are also unequal, are supposed confined to an extremely shallow layer immediately adjacent to the water surface.

Eqs. (3) and (4) can be identified with the mixing law for two moist air masses with characteristics r_0, θ_0 and $r_v(T_F)$, θ_F – the latter to be interpreted as resulting from intimate contact of air with the water surface. Then, as is well known, the result of mixing the two air masses is found in the r, θ plane on the straight line joining r_0, θ_0 and $r_v(T_F)$, θ_F – as is stated in Eqs. (3) and (4).

For fresh water the point $r_v(T_F)$, θ_F also lies on the curve of saturation mixing ratio versus temperature and reflection shows that the tangent to the curve at this point divides the r, θ plane in a significant manner. If the point r_0, θ_0 lies below the tangent, steaming of the surface cannot take place, if r_0, θ_0 lies above the tangent, steaming may take place. For in the latter case, r_0, θ_0 in the hatched area of Fig. 2, some of the air modified by mixing acquires a mixing ratio which exceeds the saturation value for its temperature; this we suppose to be a necessary condition for steaming. Thus for steaming

$$\beta < \left(\frac{dr_v}{dT}\right)_{T_F} . \quad . \quad . \quad . \quad . \quad . \quad (5)$$

and for *just no steaming*

$$\beta = \left(\frac{dr_v}{dT}\right)_{T_F} . \quad . \quad . \quad . \quad . \quad . \quad (6)$$

Given the initial condition of the cold air, Eqs. (4) and (6) permits the determination of the value of the water temperature for just no steaming. (From the graphical interpretation it is clear that Eq. (6) only possesses a solution if $d^2 r_v/dT^2 > 0$. Hutton (1788) in considering why the breath of animals is sometimes rendered visible correctly inferred that ' the solution of water in air increases with heat (temperature) in an increasing rate.' He then proceeded to a theory of clouds and rain based on mixing. Because of the observed complex dependence of r_v on T the determinations must be made numerically; results are shown in Fig. 3. The values computed are threshold values in the sense that if the air is colder or has higher relative humidity than shown, steaming can occur. From Fig. 3 it is noted that the air-water temperature difference for just no steaming increases with *increasing* water temperature and has a minimum value when the air is moist and the water cold.

The presence of dissolved inorganic salts in the ocean results in a lowering of the equilibrium vapour pressure below that for fresh water at the same temperature. If $1 - f$ is the fractional depression then $r_e = f \,.\, r_v(T)$ where r_e and $r_v(T)$ are the equilibrium values for contaminated and fresh water respectively. In the problem where deep homogeneous air overruns *saline* water of uniform temperature $T_s = \theta_s$, r and θ are again linearly related as in Eq. (3) but the slope β_s is now

$$\beta_s = \frac{f \,.\, r_v(T_s) - r_0}{\theta_s - \theta_0} \qquad \text{(7)}$$

and the condition for just no steaming is

$$\beta_s = \lambda \left(\frac{dr_v}{dT}\right)_{T_s} \text{with } \lambda < 1 \,. \qquad \text{(8)}$$

Examination of Fig. 2 should make clear the need of λ, for tangency is now required of the line joining r_0, θ_0 to r_e, θ_s with the curve of saturation mixing ratio versus temperature, where r_e no longer lies on the saturation curve. Evidently for a given condition of cold air the surface temperature of saline water necessary to promote steaming must be higher than that of fresh water and the difference is denoted as $\Delta\theta$. Using the fact that the saturation mixing ratio is accurately proportional to exp (αT) over a limited range of temperatures, it may be shown that

$$f = \{1 + \alpha\Delta\theta\} \exp - \alpha\Delta\theta \qquad \text{(9)}$$

and

$$\lambda = \exp - \alpha\Delta\theta \qquad \text{(10)}$$

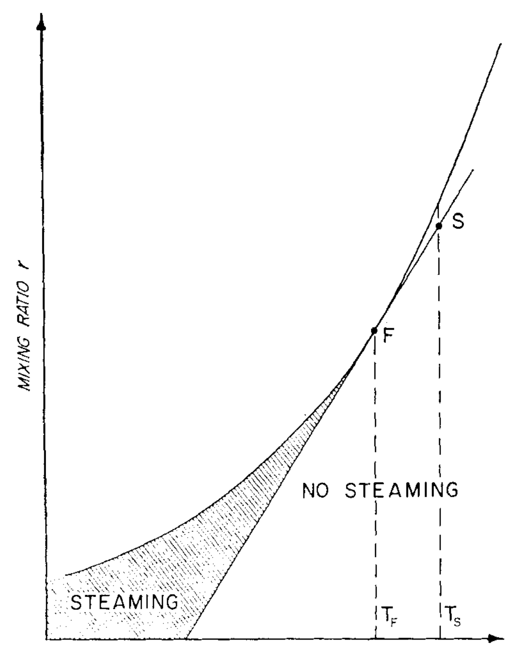

Figure 2. A criterion for the steaming of a fresh-water surface of temperature T_F and a saline-water surface of temperature T_S. SF is tangent to the saturation curve at the temperature T_F.

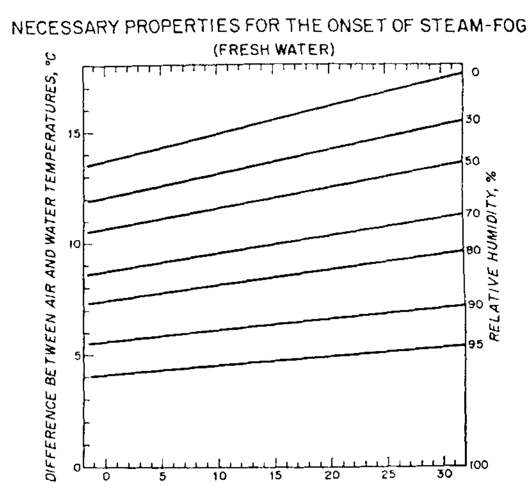

Figure 3. Necessary properties of air for the steaming of fresh water. The relative humidity is measured near the surface in the cold air.

Values of λ and $\Delta\theta$ as a function of f are shown in Fig. 4 where α has been allotted the values :

water temperature,	°C	0	15	30
α,	$°C^{-1}$	$7{\cdot}3 \times 10^{-2}$	$6{\cdot}5 \times 10^{-2}$	$5{\cdot}7 \times 10^{-2}$

The criterion for just no steaming is seen to be surprisingly sensitive to small amounts of impurities. For a depression of the equilibrium vapour pressure of only 0·01 per cent (salinity 0·2 per mille), a value reached in fresh-water lakes and rivers, $\Delta\theta$ is 0·2°C; but for a depression of 1·88 per cent (salinity 35 per mille), a value characteristic of the worlds ocean surfaces, $\Delta\theta$ is between 2·5 and 4°C ! This difference is so large that a diagram has been prepared showing the circumstances in which a saline surface is expected to steam (Fig. 5).

(b) *The liquid water content in sea smoke and steam fog*

When the characteristics of the cold air lie in the steaming region (Fig. 2), θ and r in Eq. (1) and subsequently are more logically interpreted as the wet-bulb potential temperature and the mixing ratio in both the liquid and vapour phase respectively (Rodhe 1962). However, in computing the water content here, a simpler procedure outlined by Brunt (1935) was followed.

Sample results are shown in Figs. 6 (a) and 6 (b) in conditions judged to correspond to weak and intense steaming of a saline surface. As is anticipated from the graphical interpretation of Eqs. (3) and (4) given earlier, increasing the thermal contrast between the air and water (for given r_0) increases both the water content of the fog and the *range* of air temperatures which sustain saturation. Even in extreme conditions, temperature contrast 40°C, the fog liquid water reaches a value of only about 1 g m^{-3}.

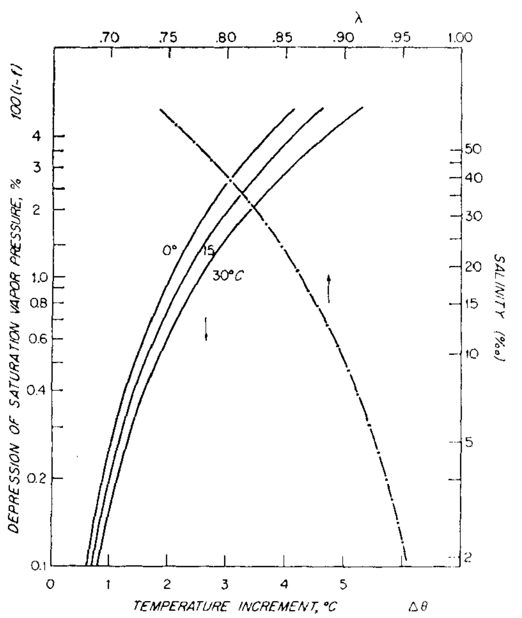

Figure 4. Difference between the surface temperature of contaminated water and of fresh water necessary to promote steaming for given cold air characteristics as a function of depression of saturated vapour pressure (solid lines) : also λ of Eq. (8) as a function of the same depression (chain line).

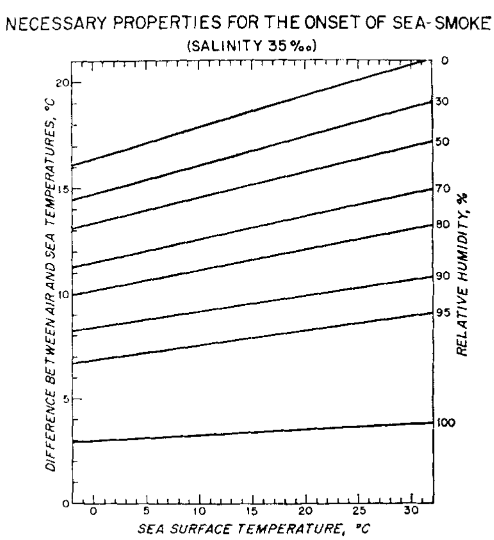

Figure 5. Necessary properties of air for the steaming of saline water (salinity 35‰). The relative humidity is measured near the surface in the cold air.

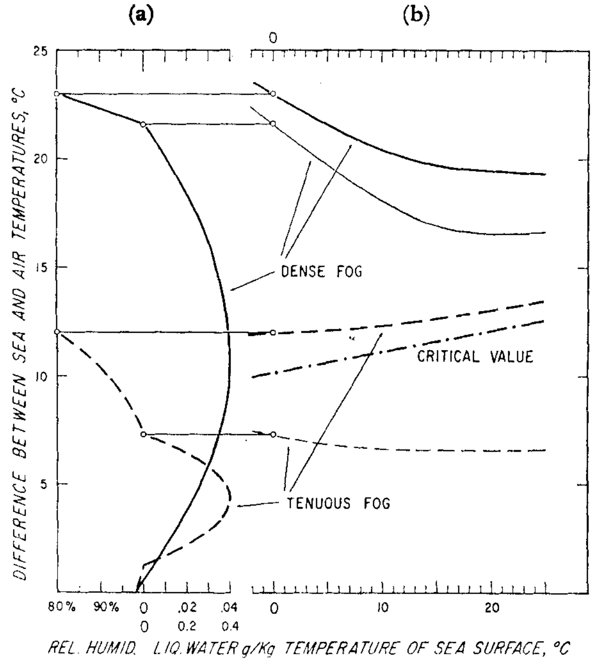

Figure 6. (a) Liquid and vapour in tenuous and dense fog; saline-water temperature 0°C and air relative humidity 80 per cent; (b) Variation of fog properties with water temperature.

4. A comparison between theory and data : conclusions

The following procedure has been adopted to test the simple ideas advanced in Section 3. Given an observation of steaming, the air temperature necessary for the onset of steaming is determined from the measured water temperature and measured relative humidity (Figs. 3 and 5); the difference between the measured and threshold air temperature is thus obtained. This difference is plotted against water temperature in Fig. 7 along with information about the vertical extent of steaming; occasions of steaming are distinguished from occasions of no steaming by the use of closed and open symbols respectively. The observations, which are drawn from reports by Church (1945), Hay (1953), *Marine Observers Log*, Rodewald (1937, 1959), Starov (1955) and Woodcock (private communication), show an approximate division between steaming and no steaming for air temperatures close to, but somewhat lower than, the theoretical values.

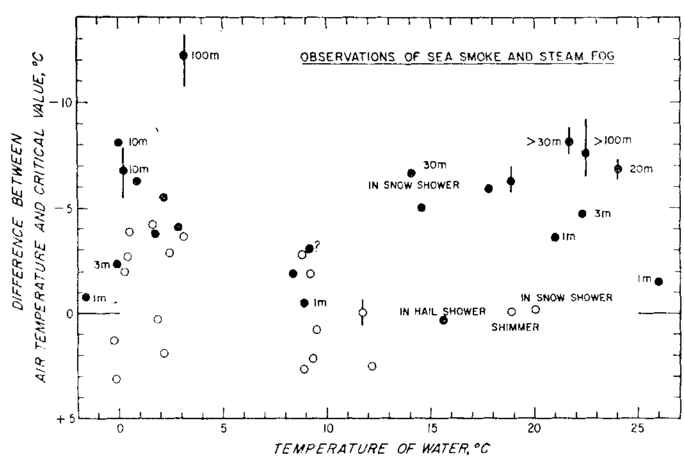

Figure 7. Reports of sea smoke and steam fog. For observers, see text, Section 4.

Careful investigation of the transition from steaming to no steaming has been made by the author on two occasions in the saline waters of Great Harbour, Woods Hole, Massachusetts. Measurements of dry- and (frozen) wet-bulb temperatures were made on the upwind shore of a peninsula in air arriving after an overwater trajectory of about 1 km : observations are presented in Table 1 and Fig. 8. On 8 February 1963 steaming persisted from patches of water until the air temperature was only 0·4°C lower than the threshold value (Fig. 8). On 21 December 1963 steaming persisted until this difference was 0·9°C.

TABLE 1. OBSERVATION OF THE TRANSITION FROM STEAMING TO NO STEAMING

(a) 8 Feb. 1963. Surface temperature − 1·7°C ± 0·1. Salinity 31·5‰. Wind 7-10 m sec^{-1} NNW. Temperature observations made at a height of 10 m

Time (LST)	Relative humidity (%)	Air temperature necessary for steaming (°C)	Measured air temperature (°C)	Difference (°C)	Steaming characteristics
0705	48·5	− 14·9	− 14·0	+ 0·9	none
0715	49·5	− 14·8	− 14·2	+ 0·6	none
0730	48	− 14·9	− 14·9	0	none
0745	45	− 15·1	− 15·0	+ 0·1	none
0810	44·5	− 15·2	− 15·1	+ 0·1	none
0815	41	− 15·4	− 15·6	− 0·2	none
0850	43·5	− 15·3	− 16·1	− 0·8	faint, widespread
0930	45·5	− 15·1	− 16·3	− 1·2	widespread, 70 cm
0950	44·5	− 15·2	− 16·4	− 1·2	1 m
1010	46·5	− 15·1	− 16·6	− 1·5	1 m
1030	43·5	− 15·3	− 16·1	− 0·8	fainter now
1050	49	− 14·9	− 15·9	− 1·0	faint, widespread
1100	45	− 15·1	− 15·7	− 0·6	in patches
1105	39·5	− 15·5	− 15·9	− 0·4	in patches
1115	43	− 15·3	− 15·6	− 0·3	in patches
1125	47·5	− 15·0	− 15·5	− 0·5	in patches
1135	48·5	− 14·9	− 15·3	− 0·4	very faint patches
1145	46	− 15·1	− 15·4	− 0·3	none
1200	46	− 15·1	− 15·1	0	none
1210	52	− 14·7	− 14·5	+ 0·2	none

(b) 21 Dec. 1963. Surface temperature 0·0 ± 0·3°C. Salinity 31·5‰. Wind 5-7 m sec^{-1} NW. Temperature observations made at a height of 2 m

Time (LST)	Relative humidity (%)	Air temperature necessary for steaming (°C)	Measured air temperature (°C)	Difference (°C)	Steaming characteristics
0945	83	− 9·6	− 12·2	− 2·6	widespread, 1 m
0955	73	− 11·0	− 12·3	− 1·3	fainter
1000	83	− 9·6	− 12·6	− 3·0	1 m
1020	85	− 9·3	− 12·8	− 3·5	
1035	83	− 9·8	− 12·0	− 2·2	50 cm
1040	83	− 9·8	− 11·7	− 1·9	50 cm
1045	77	− 10·5	− 11·7	− 1·2	faint patches
1050	81	− 9·9	− 11·0	− 1·1	none
1052	81	− 9·9	− 11·1	− 1·2	faint patches
1055	82	− 9·8	− 11·3	− 1·5	patches
1056	83	− 9·6	− 11·1	− 1·5	patches
1057	81	− 9·9	− 11·5	− 1·6	faint patches
1059	81	− 9·9	− 11·9	− 2·0	patches
1100	74	− 10·9	− 11·8	− 0·9	none
1103	80	− 10·0	− 11·2	− 1·2	none
1114	81	− 9·9	− 11·5	− 1·6	faint patches
1117	74	− 10·9	− 11·7	− 0·8	none
1120	71	− 11·2	− 11·7	− 0·5	none
1125	66	− 11·8	− 11·4	+ 0·4	none
1130	74	− 10·9	− 10·2	+ 0·7	none

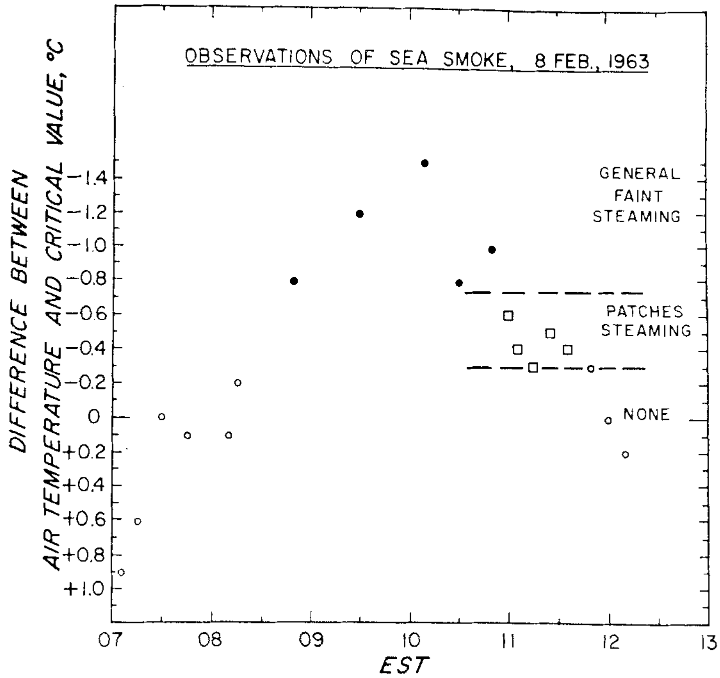

Figure 8. Observations of faint steaming, 8 February 1963 (author).

The steaming patches of sea water were presumably slightly warmer than their surroundings being produced by the upwelling induced by tidal currents and the discharge of effluents : no measurements were made of these surface temperature variations, but because of the highly stirred condition of the sea at these times it seems unlikely that they were larger than one- to three-tenths of a °C.

The data presented in the previous paragraphs indicate that, as employed to determine the conditions in which steaming occurs, the theory is an excellent first approximation. An improvement on it will need to recognize that for steaming to be apparent the conditions must have developed *beyond* the threshold stage. Thus (i) a certain minimum liquid water content (order 0·01 g m^{-3}) must be condensed out in order to provide sufficient visual contrast, and (ii) this condensed water must be raised (in turbulent fluctuations) to a certain minimum height above the water surface (order 10 cm). Condition (i) implies that for saline/fresh water the air must be approximately 0·4°C/0·7°C colder than the just no steaming value. Condition (ii) is dependent on the wind and thermal structure in the air in a way which has yet to be investigated.

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Appendix

A list of terms for the clouds of ' steam ' rising from warm water into cold air : vapour (laymen); black frost, white frost (N. Atlantic fishermen); sea mist, sea smoke, Arctic sea smoke, Arctic smoke; frost smoke and barber (crystalline.) Steam mist, autumn mist, water smoke; cold air advection fog, evaporation fog and mixing fog.

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	1957	*Ibid.*, **27**, p. 12 and 211.
	1958	*Ibid.*, **28**, p. 187.
	1960	*Ibid.*, **30**, p. 12.
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Woodcock, A. H. and Stommel, H.	1947	*J. Met.*, **4**, pp. 102-103.

* Reports of presence or absence of sea smoke used in compiling Fig. 6.

These conclusions are of paramount importance where an experimental site is a few hundred metres in extent. As a particular example, it is necessary to ensure that the slight curvature of near-neutral wind profiles is correctly apportioned between site effects and stability effects when the latter are the subject of investigation.

Dr. A. A. TOWNSEND (*in reply*) : The relation, $\partial u/\partial z = u_*/kz$ with u_* varying with height, is expected to be valid in an equilibrium layer, in which advection of turbulent energy has a negligible effect on the energy balance. In the present context, the relation is accurate in the equilibrium layer close to the surface and also at large heights where the stresses and velocity gradients still have their upstream values which are those of an equilibrium layer. The intermediate behaviour is not known and assumption of overall validity for the relation is simply a reasonable approximation. As Dr. Pasquill remarks, the predictions hardly change if u_* is assumed constant with height but both theories predict flow accelerations that require a vertical gradient of shear stress. It is logical, even if not strictly necessary, to include a stress gradient in the theory.

Taylor assumes that the surface stress becomes equal to the equilibrium value for the new surface within a very small fetch, and this implies slow and persisting acceleration of the flow except very near the surface. Both Elliott's and our work predicts overshoot of surface stress followed by a slow recovery to the new equilibrium value. I believe that our description is correct, i.e., that adjustment to the changed roughness is confined to heights less than d as defined in our paper and that adjustment of the velocity profile to the local surface stress and changed roughness is nearly complete for heights less than $\frac{1}{2}d$.

If the change of zero plane displacement produces some profile drag near the change of surface, I would expect changes in the profile over the same range of heights. For a given total drag, it should be possible to extend our method to calculate the profiles and changes of surface stress but I think that the effect of profile drag would be small for normal fetches and likely amounts of drag. Then the only difficulty is the determination of the zero plane whose position is not critical if d/z_0 is large.

Yes, I would reply to Mr. Sibbons, the interface for diffusion is nearly half the height for momentum if parallel assumptions about the shape of the distributions of flux and stress are used and if the ground conditions are such that the vertical gradient of flux is everywhere of the same sign. The difference seems to arise from the non-linearity of the equation for momentum diffusion.

I think that the 'happy accident,' mentioned by the President, depends on the nature of turbulent motion, in particular on the comparatively large values of the correlation coefficient between horizontal and vertical velocity fluctuations. Because of this, turbulent energy and Reynolds stress are equal to an order of magnitude. The energy argument could not be applied to laminar flow because the viscous stresses are usually much smaller than the energy of the molecular motion.

I agree with Dr. Taylor that the fetch necessary to ensure complete equilibrium and negligible flow accelerations is very long indeed. His example shows that details of the profiles are affected for quite small height-fetch ratios but I should point out that, while a linear variation of u_* is adequate for predicting the general form of the profile, it is a very crude approximation when it comes to calculating details such as the distribution of $kz\,(\partial u/\partial z)/u_*$. The theory provides only a criterion that the level of adjustment of friction velocity or velocity gradient is of order z/d, which is really the point made by Dr. Dyer. His comparison of the diffusive adjustment with our theory is interesting and comforting, but I should point out that the agreement between the two sets of fetches would be improved considerably by using Eq. (8) to calculate the x-d relation rather than the very crude approximation $x = 10\,d$.

551.551.8 : 551.575.1

Sea smoke and steam fog

By P. M. SAUNDERS

(Read 21 October 1964. See Q.J., **90**, p. 156)

Dr. H. L. PENMAN : I see no reason for quarrelling with the use of the linear relation between humidity and temperature. It is simply an expression of the assumed identity in shape of the profiles. I am astonished to learn that the basic physics was set out in 1788. This is about a dozen years before Dalton gave the beginnings of the concept of vapour pressure. How was it done ?

Dr. F. H. LUDLAM : I suppose that the threshold conditions for fog formation obtained by Dr. Saunders cannot be quite exact, considering the existence of a thin layer at the sea surface in which the molecular transfer coefficients for heat and moisture are not equal, and the difficulty of choosing the appropriate relation between temperature and saturation mixing ratio when a deep layer of air is involved.

Dr. J. HALLETT : If fog should form from air whose trajectory lay over an extensive snow surface far from sources of pollution, and then flowed over quiescent water, the resulting low concentration of nuclei may render any ' steaming ' almost invisible. Do you have any observations which would show that the presence or absence of fog is in any way related to the nucleus concentration ?

Dr. P. M. SAUNDERS (*in reply*) : If (i) the atmosphere were incompressible, (ii) the molecular diffusion coefficients for heat (K) and water vapour (D) were identical and (iii) the temperature and mixing ratio (strictly the number density of vapour molecules) had values which were specified along a boundary and also far from it, then there would exist *similarity* in the distributions of temperature and mixing ratio. This result, *deduced* from the diffusion equations and boundary conditions, would apply to the instantaneous as well as the average values in a turbulent flow. As Dr. Penman has pointed out, a linear relation between moisture content and temperature is an equivalent way of expressing similarity of the distributions.

Unfortunately (for our theories) the atmosphere is neither incompressible nor is the ratio K/D unity; rather $K/D = 0{\cdot}84$. Thus the threshold conditions given in the paper for fog are not exact. However, for a discussion of the onset of steaming we are concerned only with the bottom few metres of the atmosphere where the pressure variations can be neglected. The importance of the inequality of the diffusion coefficients cannot be determined on theoretical grounds, but the observations of Charnock and Ellison (1959) and those reported in this paper indicate that only quite careful field observations can be expected to reveal its magnitude.

I am aware of only one report in which the concentration of condensation nuclei had a demonstrable effect on determining the circumstances of steaming; namely the winter observations of the warm springs in Yellowstone National Park, Wyoming (Vonnegut, 1962) : an injection of nuclei into, or ionisation of the air above the springs yielded a dense steam fog where only very faint traces existed before. I suppose that similar circumstances arise in polar regions also.

Dr. Hutton's remarkable discussion of the process of condensation by mixing (1784) is developed in terms of ' the quantity of water which can be held in solution by a given quantity of air.' From the text it is evident that he identifies the properties of a mixture of water vapour and air with those of solutions of salts in water. If, as seems probable, no measurements were available to him of ' the dissolving power of air in relation to water ' then his assertion that with increasing temperature it would increase at an increasing rate, falls in a category too rare in our subject, namely a prediction subsequently verified by observation.

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551.511 : 551.511.6 : 536.2

Natural convection in water over an ice surface

By A. A. TOWNSEND

(Read 21 October 1964. See *Q.J.*, **90**, p. 248)

Mr. J. S. SAWYER (*President*) : Have any calculations been made of the rate of dissipation of gravity waves which might be stimulated in the atmosphere above inversions ? Is the degree of gravity-wave activity in the atmosphere to be regarded as a balance between generation and such dissipation, and would this provide a way of checking the reasonableness of any proposed dissipation process ?