M.O. 641 AIR MINISTRY

METEOROLOGICAL OFFICE

THE METEOROLOGICAL MAGAZINE

VOL. 87, No. 1,031, MAY 1958

ESKDALEMUIR OBSERVATORY—THE FIRST FIFTY YEARS

By M. J. BLACKWELL, M.A.

This month fifty years ago, on May 11, 1908 to be exact, it is recorded[1] that the first Superintendent of Eskdalemuir arrived to take up duties at the newly built Observatory of the National Physical Laboratory, set in the remote north-east corner of Dumfriesshire. Now, on the occasion of the golden jubilee, we can look back with not a little pride at the sequence of events and changes which have transformed an eleven acre site of wind-swept moorland into a modern geophysical observatory.

Past.—It is well known[2] that the westward extension of the London electric tramways prevented the continuance of magnetic recording at Kew Observatory, just as the electrification of the suburban railways caused the transfer of magnetic work from Abinger to Hartland, nearly fifty years later.[3] But though it was clearly desirable to establish another observatory in an undisturbed area, to carry on the work, it was probably the £10,000 compensation received from the tramway company which made the scheme a reality.

Thus, in due course, a committee of the Royal Society selected a suitable alternative site – reputedly by placing a coin on a map so as not to intersect any towns, industrial areas or railways within a radius of some ten miles. Past and present staff of the Observatory can bear witness to the thoroughness with which the committee performed their task.

Leaving behind the railhead at the Border towns of Langholm or Lockerbie, the visitor must follow the winding hill roads for fourteen miles until they meet at the small cluster of dwellings which is Eskdalemuir; here he must strike north, along the road to Ettrick and Selkirk, for a further three miles, to reach the tiny hamlet of Davington with its Kirk of Covenanter memories. A short spur road was laid up to the new site – the first metalled road in the district, and not before time, considering that Telford was born and bred in Eskdale.

We can hardly comprehend the isolation of those days: the horse-drawn transport bringing most provisions only once a month; yet such an existence made for complete acceptance of local conditions and a zest for any activities in the area. The telephone at the Observatory was the first in the district and, indeed, it was nearly decided to place the Post Office there also. Inside the enclosure, the early photographs[4] show the three buildings: Rayleigh House (Superintendent's quarters), Office Block and Schuster House (Assistants' quarters), standing out in stark relief against the barren foothills of the Southern Uplands. The names of the houses commemorate the scientists who were associated with the foundation of the Observatory.

The scientific programme of work, as it evolved over the first forty years, has already been described[5] in some detail by Mr. J. Crichton, who was Superintendent throughout the difficult years of World War II and the immediate post-war period. Apart from its original function as a geomagnetic station, the Observatory has always been concerned with meteorology, atmospheric electricity, and seismology (until 1925, when the equipment was transferred to Kew); this programme owes much to the guidance and influence of the Gassiot Committee, which administers a trust devoted to the furtherance of these subjects.

The recording of the earth's magnetic field by the original Adie magnetographs was extended, from the time of the 1932–33 International Polar Year, by the more sensitive La Cour magnetographs; at the same time, the acetylene gas supply was replaced by large storage batteries which are now used to supply recording apparatus in many parts of the establishment. A few years later, in 1936, came a further change when the Observatory was connected to the main electricity supply: thermostatically controlled electric radiators could then be used in the underground magnetograph house. A not unwelcome by-product of the change was the rapid disappearance of the oil lamps used for illumination. The basic absolute instruments, declinometer, coil-magnetometer and dip-inductor, were supplemented from about 1949 by portable sub-standards, the quartz horizontal magnetometer (Q.H.M.) and the balanced magnetometer for vertical force (B.M.Z.), for facilitating inter-observatory comparisons. For some years also, the rate of change of vertical force was detected by a large loop laid round a height contour on the moor outside the grounds.

In 1910, the Observatory came within the control of the Meteorological Office and became a first-order synoptic and climatological station. In addition to the standard equipment, a photographic barograph and thermograph, a Hellman-Fuess snow-gauge and a Jardi rate of rainfall recorder were installed. Later came the atmospheric pollution recorder, run in cooperation with the Fuel Research Station, and for a long time the night-sky recorder. In atmospheric electricity, potential gradient records have been obtained with a Dolezalek electrograph, originally using a water-dropper and later a polonium collector.

In a commemorative article such as this, we are more concerned with the broader aspects of the history and development of the Observatory; for greater detail, reference can be made elsewhere[6], and we therefore turn to the last decade of our story.

Present.—With the steady growth of the scope and detail of the work, more accommodation had to be found. Additional married quarters were built in 1928, and are now occupied by the deputy Superintendent and Mechanic; more recently, the old acetylene house was converted for use as the Handyman's Cottage. Now, in our jubilee year, the former schoolhouse at Davington has become yet another married quarter.

The stables and coach-house now house the batteries, charging room and garage. The original De Dion motor car, acquired before World War I, was followed by a wealth of makes and types of vehicle, ending in an up-to-date estate car, which brings a welcome measure of comfort to staff and families making the long journey to the nearest towns. The phonographs and early battery portables in the living quarters have given way to the latest tape-recorders, V.H.F. radios and television sets. Amidst all these changes lived the late Mr. J. B. Beck, who served from 1913 to 1950 with but a short break for war service, and Mr. W. J. Hogg, who joined the staff as Mechanic in 1920 and whose length of service is approaching the remarkable total of forty years, practically all of it spent at Eskdalemuir.

As recent photographs (between pp. 144–145) of the Observatory show, the shelter belt of Norwegian and Sitka spruce is nearing maturity; planting, which began on a large scale with a gift of sixteen hundred trees in 1911, was followed by topping and lopping of the taller trees at intervals in the 1930's. In the last few years, we have had to consider the risk of over-sheltering, particularly of the anemograph, thermometer screens, rain gauges and sunshine recorder. As a result, a co-ordinated programme of felling, thinning and re-planting was put into effect in 1956 with a view to maintaining the exposure of the instruments at a constant level in years to come. This work has led to the addition of a full time Groundsman to the staff.

The basic instrumental work in geomagnetism has changed little in recent years. Theoretical studies have been set aside so as to make more efficient use of staff arriving fresh from meteorological work elsewhere. The analysis of the magnetic records, however, becomes more involved year by year: to the daily character figures (C-figures) were added the three-hourly disturbance ranges (K-indices) and, for the International Geophysical Year, the quarter-hourly disturbance ranges (Q-indices). The magnetograms are scrutinized for sudden commencements, solar flare effects and micro-pulsations; even rudimentary attempts at forecasting magnetic storms and aurorae have been made, using the direct-vision photo-electric variometer.

There has always been a close link between the Observatories and the current scientific expedition of the day. During the International Geophysical Year, we have a particular interest in the Royal Society base at Halley Bay since all the members of the geomagnetic group received preliminary training at Eskdalemuir, and the instruments used were calibrated here. Yet, if there has been a significant trend in the scope of our geophysical work in the last decade, it has surely been in the growth of the meteorological relative to the geomagnetic aspects.

Following the extensive post-war work, in solar and terrestrial radiation, of our sister observatory at Kew, total and diffuse radiation solarimeters were installed on top of the tower stairhead in 1950; until this time, little use had been made of the tower, an ideal observational platform. Indeed, the duty observer had to climb a narrow stair and two ladders before emerging on to a square yard of icy, wind-swept roof. In 1956, the stairhead was cut away to give an uninterrupted horizon; the ladders were replaced by normal flights of stairs, and a platform of angle iron was constructed on the tower roof. To the solarimeters have been added a bimetallic actinograph and the sunshine recorder, now with a much improved exposure.

Mention has been made of our additional instruments for measuring precipitation; this branch of the work is still growing steadily: an open-scale rain recorder was installed in 1957, together with a standard Canadian snow-gauge and two experimental snow-gauges, one with a rotating head, the other with a specially designed wind-shield. For the International Geophysical Year, an American Class "A" evaporation pan was installed, and apparatus for the chemical sampling of air and rain-water.

At the outbreak of the last war, a Dobson spectrophotometer, for the determination of total atmospheric ozone content, was diverted to Eskdalemuir for several years. Now, as part of our International Geophysical Year programme, another instrument has been installed; but in addition to the standard measurements, attempts will also be made to determine the vertical distribution of ozone by means of "Umkehr" observations.

Future.—The rate of expansion in recent years has been too rapid to permit more than a hurried glance into the future. Suffice it here to mention but a few changes which are already imminent.

The staff continues to grow and, fortunately with it, so does the accommodation. The nine or ten synoptic reports of today are being increased to sixteen hourly reports, from 0600 to 2100 G.M.T., and it has been agreed in principle to extend this programme still further to cover the full 24 hours; this is only to be expected when one realizes that Eskdalemuir is almost the only inland first-order station in Scotland.

On the instrumental side, the recording of daylight illumination is being added to the radiation programme, and after that the measurement of total net flux using ventilated flux plate radiometers. An attempt to calibrate snow recorders by measuring snow depth profiles will be made. The replacement of outdated mechanical time-shutters by relays in the lamp circuits, and installation of a centralized Observatory time system has been started.

The cumbersome, photographic electrograph will be replaced by a valve-voltmeter circuit and recorder; less immediate, but nevertheless desirable, is the use of the new nuclear-precession magnetometer for inter-Observatory comparisons. Lastly, with radiation, evaporation, wind and temperature records available, we are well under way towards the setting up of an ideal hydrological experiment, for which nature has provided us with the nearby Davington burn and its well defined catchment area. So—here's to the next fifty years! May we be worthy of those who led the way.

REFERENCES

1. London, National Physical Laboratory, Observatory Dept.; Report for the year 1908. London, 1909, p. 14.
2. London, Meteorological Office; Observatories' Year Book 1938. London, 1955, p. 84.
3. London, Royal Astronomical Society; Report of the Astronomer Royal for the year 1954–55. London, 1955, p. 4.
4. London, Meteorological Committee; Report for the year 1910–11. London, 1911, facing p. 48.
5. CRICHTON, J.; Eskdalemuir Observatory. *Met. Mag., London,* **79**, 1950, p. 337.
6. London, Meteorological Office; Annual Reports of the Director of the Meteorological Office. London.

HEAT SOURCES AND SINKS AT THE EARTH'S SURFACE

By D. M. HOUGHTON, M.Sc., D.I.C.

Summary.—A study has been made of the spatial and temporal variations in the surface character of the earth which can result in increases or decreases in

the heat supply to the atmosphere. For the sea areas the changes in surface temperature and in the ice limits are considered and for the land the changes in albedo, evaporation and evapotranspiration, state of surface, heat-storage capacity and conductivity are dealt with. Some of these factors are found to be unimportant, while others can exert an appreciable and sometimes long-term influence upon the heat supply to the atmosphere.

Introduction.—The solar radiation received at the outside of the atmosphere is 800 to 900 calories per square centimetre per day throughout the year at the equator and at the poles it is 1070 cal. $cm.^{-2}$ day^{-1} in midsummer, falling to nil in winter. If we take a long period average covering the northern hemisphere then approximately 47 per cent of this radiation reaches the earth's surface and about 19 per cent is directly absorbed by the atmosphere and clouds.[1] Some of the radiation received at the earth's surface is reflected back to space, some is absorbed and stored and some is passed to the atmosphere by convection and radiation. All of these processes are affected by the physical character of the surface, with the result that the amount of heat reaching the atmosphere either immediately or by the subsequent transfer of stored heat from the surface may be significantly affected. Here we are concerned with those effects which may be sufficient to affect the circulation of the atmosphere over a substantial area for a period of days. Such effects must occur over a substantial area and persist for a few days at least. We therefore limit the study to changes which can affect an area of a minimum of 10^4 square kilometres for 5 days at least.

The magnitudes of the variations in heat supply to the atmosphere may be compared with the outgoing long-wave radiation to space (500 cal. $cm.^{-2}$ day^{-1} in low latitudes, 380 cal. $cm.^{-2}$ day^{-1} in high latitudes), or with the change in the heat reflected to space (and consequently lost to the atmosphere) due to a change in cloudiness from nil to 8 oktas (100 to 200 cal. $cm.^{-2}$ day^{-1} in low latitudes). However, it is probably safer to compare the variations with one another and see which are the most important.

The energy of the atmospheric circulation is derived primarily from horizontal differences of temperature. These horizontal temperature gradients may be affected by variation in the supply of either sensible or latent heat to the atmosphere. There is however the important difference that the modification to the temperature field resulting from variations in supply of latent heat will take place when condensation occurs and this may be at a distance of several hundred kilometres from the source of evaporation.

The Oceans.—Of all the heat stored and released by the earth's surface during the course of the annual temperature cycle 86 per cent is stored by the oceans.[2] Most of this is released in winter months and at a rate which depends upon the temperature difference between the sea and the air above. As far as the sea itself is concerned the determining factors in its heat output are its surface temperature and whether or not it is frozen.

Sea surface temperature.—When the sea surface is warmer than the air above it, heat is transferred upwards by convection and long-wave radiation. Convective processes are also set up in the sea itself so that the sea surface temperature changes little.

When the sea surface is colder than the air above it, stable conditions are established in both media and heat is transferred downwards by long-wave radiation and turbulent diffusion. Cloudy conditions are likely and there will be net downward radiation from the cloud base, but even for a cloud–sea temperature difference of 5°C. (which is rare) this is only about 25 cal. $cm.^{-2}$ day^{-1}.

A relationship between the flux of heat and water vapour to the atmosphere and the air–sea temperature difference has been suggested by Jacobs[3] but is applicable only to mean values over a period of about a season. Special cases covering shorter periods have also been analysed from time to time and provide some indication of the variation in intensity of convective heat transfer over the oceans.

(i) The long-term heat budget:

Jacobs gives the empirical relation

$$Q_e = 0\cdot143\ (e_w - e_a)W_a$$

where Q_e = heat equivalent of evaporated water (calories)

W_a = wind speed (metres per second)

e_w = saturation vapour pressure at sea surface temperature (millibars)

e_a = vapour pressure of the air at deck level (millibars).

The associated transfer of sensible heat (Q_c) is then found using Bowen's ratio. No distinction is made between rough and smooth conditions. Jacobs' charts showing the mean seasonal values of Q_c and Q_e over the North Pacific and North Atlantic Oceans give winter values for Q_c of 20 to 80 cal. $cm.^{-2}$ day^{-1} to the west of Ireland rising to over 200 cal. $cm.^{-2}$ day^{-1} off the east coast of North America. The annual means for Q_c are 20 and 100 cal. $cm.^{-2}$ day^{-1} respectively. Corresponding values for Q_e are about $2\frac{1}{2}$ times Q_c and the annual means for (Q_c+Q_e) rise from about 70 cal. $cm.^{-2}$ day^{-1} off western Ireland to 350 cal. $cm.^{-2}$ day^{-1} off eastern America.

The observed annual amplitudes of sea surface temperature variation at the weather ship stations "I" and "J" support these figures. The mean amplitude through the top 100 metres is about 4·5°C., so that the heat given up by the sea over the six winter months is about 45,000 cal. $cm.^{-2}$, an average of about 250 cal. $cm.^{-2}$ day^{-1}. Of this about half is accounted for by long-wave radiation. Sverdrup[4] quotes similar figures for the Bay of Biscay. Over these areas of the eastern Atlantic the annual mean of the air–sea temperature difference is between −1°C. and −2°C.

(ii) Shallow convection:

Riehl[5] has studied the atmospheric heat balance over a period of a few days in the trade wind regions of the North Pacific Ocean. His data satisfy the empirical equation

$$Q_c = 0\cdot036\ (T_w - T_a)W_a,$$

with values of Q_c ranging from 10 to 30 cal. $cm.^{-2}$ day^{-1} for an air–sea temperature difference of 0·5 to 1·0°C.

Data for the Baltic are contained in a paper by Hela[6] who studied the heat exchange at *Finngrundet* (a lightship). He gives the mean input of sensible heat to the atmosphere during winter months as about 50 cal. cm.$^{-2}$ day^{-1}. It is probable that shallow convection obtained on the majority of occasions, so that this figure is applicable here. He suggests the equation

$$Q_c = 0{\cdot}032\ (W_a + 0{\cdot}3)^{\frac{1}{2}}(T_w - T_a)\theta_w$$

to fit his data, where θ_w is the sea surface temperature in degrees absolute.

For cases of shallow convection over the North Sea but with a much larger temperature difference (about 10°C.) the author has found the sensible heat input to be about 40 to 80 cal. cm.$^{-2}$ day^{-1}. The data used were for very cold easterlies circulating around a Scandinavian anticyclone.

(iii) Strong convection:

Craddock's[7] paper on the modification of polar airstreams contains useful data. He shows that the sensible heat input to the atmosphere can reach values as high as 1400 cal. cm.$^{-2}$ day^{-1}, while 300–500 cal. cm.$^{-2}$ day^{-1} is probably quite normal in direct airstreams from the north. These figures are not strictly comparable with those already given, as an unknown proportion of the observed heating is due to release of latent heat in the shower precipitation during the passage of the air over the sea. The equations

$$Q_c = 0{\cdot}064\ (T_w - T_a), \text{ and}$$
$$Q_e = 0{\cdot}017\ L\ (T_w - T_a)$$

fit his graphs relating the sensible heat and water vapour transfer to the air–sea temperature difference.

A similar type of air mass is studied by Burbidge.[8] Using his data for the modification of continental polar air crossing the Hudson Bay in the late autumn the input of sensible heat is at least 344 cal. cm.$^{-2}$ day^{-1}. No account has been taken by Craddock or Burbidge of the long-wave radiational cooling of the atmosphere for which the relevant figure is of the order of 150 cal. cm.$^{-2}$ day^{-1}.

Winston[9] relates intense cyclogenesis in the Gulf of Alaska with areas of rapid heating from the sea surface. He finds the non-adiabatic component of heating in the five cases of cold air masses studied to lie between 400 and 2210 cal. cm.$^{-2}$ day^{-1}. Unfortunately values of sea and air temperature are not given.

(iv) Estimation of the effect of an anomaly of 1°C. in sea temperature on the heat transport to the atmosphere:

Several relationships have been suggested which give some indication of the dependence of the supply of sensible and latent heat to the atmosphere upon the sea surface temperature. For the mean effect over a period of months the partly empirical equations derived by Jacobs and Hela indicate that an anomaly of ± 1°C. in air–sea temperature difference will give ± 30 cal. cm.$^{-2}$ day^{-1} in sensible heat, and ± 60 cal. cm.$^{-2}$ day^{-1} in latent heat available to the atmosphere. A relationship based on Craddock's analysis suggests values up to ± 80 cal. cm.$^{-2}$ day^{-1} for the increase in the supply of sensible heat corresponding to ± 1°C. in strongly convective situations, while the latent heat supply will change by ± 160 cal. cm.$^{-2}$ day^{-1}.

Now a change in the temperature of the sea surface is always accompanied by a change in the temperature of the air immediately above it. It is reasonable to suppose that in the case of off-shore advection of comparable air masses over sea areas close to continents an anomaly of 1 °C. in sea surface temperature will correspond to an anomaly of 1 °C. in air–sea temperature difference. The above figures can then be applied as they stand and they may be assumed to indicate the maximum effect on the heat input to the atmosphere of a 1 °C. sea temperature anomaly. In all other cases the air–sea temperature difference anomaly will be some fraction of the sea temperature anomaly. An independent estimate of the size of this fraction may be obtained by considering the change in heat storage capacity of the sea itself. Anomalies in sea surface temperature are sometimes observed to last for several months. If, as is reasonable, a positive anomaly of 1 °C. extends to 50 metres depth and disappears after 5 months then the atmosphere has received an additional 50 cal. cm.$^{-2}$ day^{-1} (corrected for approximately 5 cal. cm.$^{-2}$ day^{-1} lost as additional long-wave radiation). This suggests that the mean effect of a 1 °C. anomaly is appreciable over quite a long period and somewhere near half the maximum effect indicated by the empirical equations.

Ice.—

(i) Freezing:

When the sea freezes the transfer of heat from the sea to the air is very much reduced. This is demonstrated most markedly in observations of the modification of polar continental air over Hudson Bay before and after freezing.[8] The daily input of heat to the air crossing the Bay before the surface freezes is at least 344 cal. cm.$^{-2}$ (see p. 135). After the Bay freezes in December hardly any modification can be observed.

(ii) Conduction:

If the temperature of the water beneath the ice is 0°C. and the air temperature well below this value, the rate of heat transport through the ice may be estimated for various thicknesses of ice. For ice 1 metre thick and a temperature difference across the ice of 10°C., assuming a thermal conductivity of 0·005 cal. (sec. cm. °C.)$^{-1}$, it is about 20 cal. cm.$^{-2}$ day^{-1}. In areas where the sea freezes the atmosphere loses heat by long-wave radiation at the rate of about 380 cal. cm.$^{-2}$ day^{-1}. Hence any heat received by conduction through the ice is negligible in comparison with the loss of heat by radiation.

(iii) Albedo:

Changes in the albedo of the sea through freezing need not be considered, as the incident solar radiation in the areas concerned is either nil or too small to matter.

Ocean currents.—The subject of ocean currents is beyond the scope of this paper, but it must be noted that it is these currents which largely determine the latitudinal and longitudinal distribution of sea temperatures. A sea temperature anomaly is normally due to either an anomaly in absorbed radiation or an anomaly in the direction or rate of flow of ocean currents. Also, as a result of ocean currents the areas of anomalous radiation and resulting anomalous sea temperature may become far removed from one another. The time lapse is important. For instance a large sea temperature anomaly in the Caribbean Sea and off the coast of Florida may be expected to be transferred north-eastwards in the Gulf Stream and to reach the north-eastern Atlantic in the course of a year. The maximum flow which is found off Chesapeake Bay is about 3 knots, while over most of its course the Gulf Stream transports water at about $\frac{1}{3}$ to $\frac{1}{2}$ knot.[4]

Heat exchange between atmosphere and ocean—summary.—The exchange of heat (sensible and latent) between the sea and the atmosphere is negligible when the sea is frozen, and small ($\sim$ 20 cal. cm.$^{-2}$ day^{-1}) when it is colder than the air. It is significant ($\sim$ 50 cal. cm.$^{-2}$ day^{-1}) when mechanical turbulence and shallow convection pass heat to the air and it is very large (up to 1,200 cal. cm.$^{-2}$ day^{-1}) in strongly convective situations. Downward long-wave radiation may supplement the small turbulent heat flux when the sea is colder than the air but is normally much smaller than 25 cal. cm.$^{-2}$ day^{-1}.

The maximum change in heat supply to the atmosphere over the open ocean resulting from an anomaly of ± 1°C. in sea surface temperature is estimated at ± 15 cal. cm.$^{-2}$ day^{-1} for sensible heat and ± 45 cal. cm.$^{-2}$ day^{-1} when latent heat is included. This applies to conditions averaged over a period of months. The latter figure is confirmed by an estimate based upon maximum observed sea temperature changes. In strongly convective situations these values may be increased to ± 40 and ± 120 cal. cm.$^{-2}$ day^{-1} respectively.

Observed sea temperature anomalies of the order of 1 to 3°C. and the corresponding increase or decrease in energy made available to the atmosphere must have some effect on subsequent synoptic developments.

The freezing of the sea surface, although it has been shown to be important, only occurs on a comparatively small scale. The freezing of large areas of partially enclosed seas which is an annual occurrence in high latitudes is most important in connexion with the modification of local air masses. It is possible that the late freezing of areas such as Hudson Bay might also hinder the establishment of the semi-permanent continental cold troughs in the troposphere, and the abnormal freezing of any such area should also have an observable effect on the subsequent thickness pattern in the troposphere. However, there are few areas in which the dates of freezing of substantial areas of the sea have significant variations. For instance the Baltic Sea is subject to a certain amount of freezing every winter, but it seems improbable that a variation of 20 to 50 miles in the freezing limit can exert any appreciable large scale control on the atmospheric circulation.

The Continents.—The non-uniformity of the land surface and its comparative inability to store and conduct heat necessitate a different approach to the question of whether any changes in its character can appreciably change the heat supply to the atmosphere.

Little of the heat which is received by a land surface from the sun is stored in the ground for more than 24 hours. Some of it is returned to space by reflection and radiation, some passes to the atmosphere by long-wave radiation or convection and some is used in evaporation, in the melting of snow, etc. Any increase in the amount of heat reflected or in the amount used in evaporation will reduce the amount of sensible heat made available to the atmosphere and it is therefore convenient to think of the increase of heat dissipation by any such process as a potential local loss to the atmosphere.

Heat conduction and storage.—

(i) Unfrozen ground:

Heat flow to and from the surface depends upon the thermal conductivity of the materials in the surface layers, on the seepage of water downwards associated with rainfall and on the temperature gradients which are established primarily as a result of incoming and outgoing radiation. All these are very variable but the purpose of this study will be achieved if extreme values can be estimated.

Using known values of thermometric conductivity (K) density (ρ) and thermal capacity (c), the following values of heat flux are calculated assuming a temperature gradient of 10°C. per metre (Table I).

TABLE I—VALUES OF HEAT FLUX CALCULATED FROM THERMOMETRIC CONDUCTIVITY (K) DENSITY (ρ) AND THERMAL CAPACITY (c)

	ρc	$K \times 10^{-2}$	Flux
			cal. cm.$^{-2}$ day^{-1}
Wet sand	0·53	0·60	28
Dry sand	0·30	0·16	4
Dry loam	0·4	0·2	7
Marsh	1·00	0·12	10
Clay	1·4	0·2	24
Rock	0·5	1·2	52
Snow	0·12	0·2	2
Ice	0·45	0·78	30

Temperature gradients of the order of 10°C. per metre are only found in conditions of strong insolation, or radiation under clear skies at night, but for short periods they may be considerably greater in the top few centimetres. On a hot summer's day at Dresden, Schreiber[10] found the maximum midday heat flux into the sandy soil to be about 70 cal. cm.$^{-2}$ day^{-1}. Roach[11] using a flux plate at 6 centimetres depth at Kew measured a maximum flux of 5 milliwatts (100 cal. cm.$^{-2}$ day^{-1}) which lasted for about 3 hours on a hot June day. With the same instrument he measured a maximum upward flux of 1 milliwatt during a very cold easterly outbreak in January 1954. This flux persisted for at least 24 hours. Confirmatory evidence is obtained by examining earth temperature records for Berlin and Dublin which show that the maximum 24-hour temperature changes recorded with warm air flowing over cold land or vice versa are about 3°C. in the top half metre which correspond to a maximum heat flux of 30 cal. cm.$^{-2}$ day^{-1}. For a long period average the total change in the heat content of the soil at Berlin between October and March gives a value of about 1,700 cal. cm.$^{-2}$. This may be compared with the corresponding figure for the eastern North Atlantic which is about 45,000 cal. cm.$^{-2}$ (see Sea surface temperature). Thus we see that the modification of air masses by sensible heat stored in the ground is not an important process in the general circulation.

(ii) Frozen ground:

Normally the ground is moist before winter freezing takes place and the soil is maintained for a period at 0°C. while heat is abstracted from or communicated to it during freezing or thawing. Higher temperature gradients than are normally observed may thus be established above a frozen layer of soil and may persist for considerable periods.

Let us consider a square centimetre of earth 10 centimetres in depth which contains 4 grammes of liquid water. The heat liberated during the freezing process is about 320 calories and the normal time taken to freeze to this depth is 4 to 10 days, giving a maximum heat flux over 4 days of 80 cal. $cm.^{-2}$ day^{-1}. Thawing may proceed rather more rapidly as solar heating of the soil surface gives higher temperature gradients. At Berlin 10 centimetres are observed to thaw in as little as 2 days and 25 centimetres in 7 days. The heat absorbed is a maximum of about 150 cal. $cm.^{-2}$ day^{-1}, some of which is directly removed from the atmosphere and the rest would otherwise have been transmitted to the atmosphere by convection from the heated ground. Thawing by the advection of a warm cloudy airstream proceeds at about half this rate and the maximum heat lost from the atmosphere by radiation and diffusion is about 80 cal. $cm.^{-2}$ day^{-1}.

These fairly high values of heat flux are important because of the length of time for which they may persist. They suggest that frozen ground could prevent normal atmospheric heating to a significant extent, so that variations in the semi-permanent limits of frozen ground could be important.

Radiative properties.—Albedo.—Reasonably accurate measurements of albedo are available for most types of vegetation and bare earth surfaces, though the dryness of the vegetation or earth can give quite wide variations. The following values are to be found in other publications.[1, 4, 12, 13, 14, 15]

TABLE II—PERCENTAGE OF INCIDENT RADIATION WHICH IS REFLECTED

	%
Forest	3–10
Snow covered forest	10–25
Grass fields dry	20–30
Grass fields wet	8–20
Bare ground dry	10–25
Bare ground wet	8–10
Sand	20–25
Rock	12–15
Snow new	70–80
Snow old	55
Water	5–50 (depending on solar elevation)
Desert	24–30

In order to estimate the maximum effect of a change in albedo on the heat supply to the atmosphere it is reasonable to assume that all the radiation absorbed at the ground is redistributed to the atmosphere during 24 hours.

The largest albedo change is from a normal grass or earth surface (15 to 20 per cent) to a snow covered surface (70 to 80 per cent). Much winter snowfall occurs when the incoming solar radiation is so small that a change in albedo is insignificant in its effect on the heat supply to the atmosphere. However some areas as far south as 50°N., notably the southern shores of Hudson Bay[16] and parts of central Siberia receive their first snowfall towards the end of September, when the mean sun and sky radiation is 180 cal. $cm.^{-2}$ day^{-1}.[17] In this case the 60 per cent increase in albedo consequent on snowfall will reduce the heat supplied to the atmosphere by 108 cal. $cm.^{-2}$ day^{-1}.

These same areas are the last to lose their snow cover in such southern latitudes and substantial areas of snow covered ground may persist until late April when the mean incident sun and sky radiation is as much as 300 cal. cm.$^{-2}$ day^{-1}; 180 cal. cm.$^{-2}$ day^{-1} is then denied to the atmosphere. Between these extremes the importance of snow cover decreases with increasing latitude and increasing proximity to the winter solstice. Snow cover in Scandinavia for instance will only be important before the end of October or after the end of February.

Of other possible changes in albedo the most important is from 15 per cent with green grass to 25 per cent with parched earth. This may occur in areas where the incident radiation is 500 cal. cm.$^{-2}$ day^{-1}, when the 10 per cent change in surface reflectivity will increase or decrease the heat available to the atmosphere by 50 cal. cm.$^{-2}$ day^{-1}.

Snow.—The change in albedo due to snowfall has already been discussed. There remains the effect of the absorption of latent heat on melting. A maximum rate of melting is about 5 centimetres of snow per day (about 0·5 grammes) and the corresponding absorption of latent heat is 40 cal. cm.$^{-2}$. As in the case of thawing earth this is a loss to the atmosphere.

Evaporation and evapotranspiration.—The thermal effects on a land surface as a result of evaporation and evapotranspiration are substantially different from the corresponding effects on a sea surface. By means of convection in the sea surface the temperature remains almost constant throughout the evaporation process. Over land the latent heat of evaporation must be supplied either by the sun, by the atmosphere or by conduction through the surface. As we have seen, the heat made available by conduction from that stored in the soil is negligible and the heat required for evaporation can be regarded either as a direct loss to the atmosphere or deducted from the solar energy which would otherwise have been communicated to the air by convection.

Numerous measurements indicate that the normal daily evaporation varies from 0·6 centimetres per day from a moist surface in the tropics to 0·05 centimetres per day from a dry surface in high latitudes. Table III gives a selection of values from the literature for evaporation and evapotranspiration.

TABLE III—A SELECTION OF VALUES FOR EVAPORATION AND EVAPOTRANSPIRATION AS GIVEN IN OTHER PUBLICATIONS

Surface and vegetation	Evaporation	Source (see References)
	cm. per day	
Grass, Toronto, November	0·05	18
Very dry soil, California	0·06	19
Bare soil, Toronto, November	0·07	18
Dry soil with vegetation, Florida	0·2	18
Bare soil, Toronto, July	0·4	18
Moist soil with vegetation, Florida	0·5	18
Grass, Toronto, July	0·5	18
Moist sand, California	0·63	18
Grass, British Isles (annual mean)	0·14	20
	cm. per year	
Savanna, tropics	92·5	18
Very wet rain forest	152·9	18

If all the latent heat of evaporation is considered to be a loss for the atmosphere then the deficit can be as much as 260 cal. cm.$^{-2}$ day^{-1}. This deficit will vary according to the vegetation or crops growing and the moisture available. After a prolonged drought the heat available to the atmosphere may be as much as 200 cal. cm.$^{-2}$ day^{-1} more than after a prolonged wet spell. This is partly offset by the increase in the albedo; but even so the net result should be evident in a much greater heating of the atmosphere over deserts in subtropical regions than over areas covered in green vegetation. The temporary formation of a new desert or vegetation over a normally desert area should be significant.

It has been assumed in the foregoing paragraphs that the heat used in evaporation is not returned to the atmosphere by re-condensation in the same region. If showers occur this will not be true nor if the evaporation is balanced by deposition of dew. However in many circumstances the air will be transported hundreds of kilometres before the latent heat is released in some region of generally ascending air such as the frontal system of a depression. Under these circumstances any variation in the evaporation in a particular area may be regarded as providing an abnormal heat sink for the atmosphere in the area of evaporation and a corresponding abnormal heat source in some other region. The net world-wide effect over a period will always be nil.

Summary.—The probable maximum effects of variations in land surface conditions on the heat supply to the atmosphere are as follows:—

TABLE IV—THE PROBABLE MAXIMUM EFFECTS OF VARIATIONS IN LAND SURFACE CONDITIONS ON THE HEAT SUPPLY TO THE ATMOSPHERE

Method	Source	Magnitude
		cal. cm.$^{-2}$ day^{-1}
Conduction	Calculated—wet sand; temperature gradient 10°C. per metre	±28
	Observed—Berlin; glacial sand and gravel... ...	±30
Freezing	Latent heat liberated: calculated: water content 0·4 gm. per c.c.	+80
Thawing ground ... (sunny conditions)	Latent heat absorbed: same water content	−150
Thawing ground ... (cloudy conditions)	Latent heat absorbed: same water content	− 80
Thawing snow ...	Latent heat absorbed in melting 5 cm. per day ...	− 40
Albedo	Change due to snow: middle latitudes late autumn and early spring	−180
	Change due to drought on green vegetation	− 50
Evaporation	Possible change due to drought or vegetation on desert	±200

Conclusion.—Before reviewing the ability of the earth's surface to modify through various means the working of the atmospheric heat engine, mention must be made of the solar constant. For many years it has been suggested that variations in the solar constant can give rise to large scale climatic changes and that small variations due to sunspot cycles are responsible for shorter fluctuations in world weather. Krivsky[21] purports to find a definite relationship between sunspot cycles and the position of the jet stream. The most useful summary of the present state of this subject is by Wexler.[22] Although he is probably more disposed than many to advocate a relationship between sunspot cycles and the subsequent weather he states that there is as yet no proof of one and, what is more, that no proof can be readily produced. Correlations between random factors have been found which are equal to those found between weather phenomena and sunspot cycles.

The foregoing survey has covered the principal ways in which variations in the state of the earth's surface can change the energy available to the atmosphere. The following appear to be the most important:

(i) an anomaly in sea surface temperature

(ii) variations in the limits of frozen land and sea

(iii) early and late snowfall in middle latitudes

(iv) change from moist land or land covered in green vegetation to desert and vice versa.

A natural sequel to the present investigation would be to seek evidence in the behaviour of the atmosphere of the effect on subsequent weather of anomalies of surface conditions of the kind listed above. The effects of the limits of arctic ice have already been considered by Schell[23] and Walker[24] but the relations with subsequent weather have been found to be weak. Possibly this might be expected from the very limited variations in the boundary of arctic ice.

A summary of other attempts to find associations between sea surface temperature, sea ice anomalies and subsequent weather has been made by Baur.[25] In particular Sandstrom suggested that a warm Gulf Stream should result in a warm winter over Europe. However the winter which followed a peak in the Gulf Stream in 1939 was the coldest in Europe for 110 years and the above-normal sea temperatures of 1955 were also followed by an abnormally cold February. Thus the search for such relations will not be easy and the effect of anomalous heat sources and sinks may be found more easily in anomalies of the atmospheric circulation over a wide area rather than in direct effects on local temperature.

REFERENCES

1. HOUGHTON, H. G.; On the annual heat balance of the Northern Hemisphere. *J. Met., Boston,* **11,** 1954, p. 1.
2. SCHUMANN, T. E. W. and BURDECKI, F.; The seasonal storage of heat in the surface layers of the earth. Pretoria, 1953. Notos **2,** p. 4.
3. JACOBS, W. C.; On the energy exchange between sea and atmosphere. *J. Mar. Res., New Haven,* **5,** 1942, p. 37.
4. SVERDRUP, H. U.; The Oceans. New York, 1940.
5. RIEHL, H., YEH, T. C., MALKUS, J. S. and LA SEUR, N. E.; The northeast trade of the Pacific Ocean. *Quart. J. R. met. Soc., London,* **77,** 1951, p. 334.
6. HELA, I.; On the energy exchange between sea and atmosphere in the Baltic. *Mitt. met. Inst. Univ. Helsingf., Helsinki,* 1951, No. 70.
7. CRADDOCK, J. M.; The warming of Arctic air masses over the eastern North Atlantic. *Quart. J. R. met. Soc., London,* **77,** 1951, p. 355.
8. BURBIDGE, F. E.; The modification of continental polar air over Hudson Bay. *Quart. J. R. met. Soc., London,* **77,** 1951, p. 365.
9. WINSTON, J. S.; Physical aspects of rapid cyclogenesis in the Gulf of Alaska. *Tellus, Stockholm,* **7,** 1955, p. 481.
10. SCHREIBER, P.; Micrometeorology. See Sutton (15).
11. ROACH, W. T.; Measurements of atmospheric radiation and the heat balance at the ground at Kew. *Met. Res. Pap., London,* No. 936, 1955.
12. BUDYKO, M. I., BERLYAND, T. G. and SUBENOK, L. I.; Heat balance of the surface of the earth, Leningrad 1954. Izvestiya, Akademii Nauk U.S.S.R. Seriya Geograficheskaya, No. 3.
13. FRITZ, S.; The albedo of the ground and atmosphere. *Bull. Amer. met. Soc., Boston,* **29,** 1948, p. 303.
14. Smithsonian Institution Meteorological Tables. Washington, D.C. 1952.
15. SUTTON, O. G.; Micrometeorology. New York, 1953.
16. THOMSON, A.; The climate of the Hudson Bay region. *Trans. roy. Soc., Can., Ottawa,* **48,** 1954, Ser. III, p. 1.
17. BLACK, J. N.; The distribution of solar radiation over the earth's surface. *Arch. Met., Hamburg,* **7,** 1956, Ser. B., p. 165.
18. MATHER, J. R.; Micrometeorology of the surface layers of the atmosphere. Johns Hopkins University, Publications in Climatology, **7,** Seabrook, N.J., 1954.

19. VEIHMEYER, F. J. and BROOKS, F. A.; Measurements of cumulative evaporation from bare soil *Trans. Amer. Geophys. Un., Washington*, **35**, 1954, p. 601.
20. PENMAN, H. L.; Survey of meteorology in agriculture. *Quart. J. R. met. Soc., London*, **75**, 1949, p. 293.
21. KRIVSKY, V.; Variation of the frontal zone in relation to solar activity. Czech. Acad. Sci., Astrophysical Obsns. Pub. No. 23, Prague, 1953.
22. WEXLER, H.; Radiation balance in climatic change. Climatic change. Edited by Harlow Shapley, Cambridge, 1953.
23. SCHELL, I. I.; Polar ice as a factor in seasonal weather. *Mon. Weath. Rev., Washington*, Supplement No. 39, 1940, p. 27.
24. WALKER, SIR GILBERT; Arctic conditions and world weather. *Quart. J. R. met. Soc., London*, **73**, 1947, p. 226.
25. BAUR, F.; Extended range weather forecasting. Compendium of Meteorology. Boston, 1951, p. 814.

VARIATION WITH TIME AND DISTANCE OF HIGH-LEVEL WINDS OVER MALAYA

By L. S. CLARKSON, M.Sc.

Summary.—Based mainly on an analysis of the statistical parameters characterizing the distribution and variation with time and distance of radar-observed winds at 40,000 feet and 50,000 feet over Singapore and Songkhla, data is deduced from which future winds at places in Malaya may be predicted by means of a simple regression on the latest Singapore observation.

Introduction.—Within about 10° latitude of the Equator, where geostrophic control is lacking, no relationship between wind and pressure gradient has been found suitable for practical application[1]. In the Malayan area there are insufficient wind observations at high altitudes to permit instantaneous stream-lines and isotachs to be drawn objectively in accordance with the methods developed by Bjerknes[2] and Sandström[3] and applied to the tropics by Palmer[4]. Furthermore, no synoptic dynamical principles have been established by means of which the future wind flow may be anticipated from that depicted on a current stream-line chart. In these circumstances the statistical methods developed by Durst[5], and shown by Johnson[6] to yield results comparable in accuracy with the orthodox forecasting techniques applicable in temperate latitudes, are applied to the practical problem of high-altitude wind prediction for aviation in Malaya.

Observations.—The only observations of radar winds at 40,000 feet and 50,000 feet in the Malayan area from which the statistical parameters in the regression equation expressing the most probable variation of wind with time and distance may be evaluated are:—

(i) the Singapore observations analysed earlier[7], but including all data for 1955,

(ii) observations from Songkhla at 0300 G.M.T. kindly supplied by the Director, Meteorological Department of the Royal Thai Navy. Songkhla is 390 nautical miles north-north-west of Singapore.

Tables I and II give the mean monthly components from the north and east, V_N and V_E, mean scalar and vector winds, V_S and $\mathbf{V}$, standard vector deviation, σ, and constancy, α, derived from the Singapore and Songhkla observations at 40,000 feet and 50,000 feet respectively.

As is to be expected, the Songkhla winds show similar general features to those found for Singapore[7]. However, mean easterly components are appreciably less at Songkhla than at Singapore in the months from November through to April; at neither place is there much evidence for the winter maximum referred to by Hay[8] and Gilchrist[9]. At 50,000 feet the sharp increase in easterly component from April to May is especially pronounced: the strong summer east-north-easterlies at Songhkla are as remarkably steady in direction as those at Singapore, with a constancy exceeding 90 per cent, but show a rather greater variability about the vector mean.

The data in Tables I and II point to the onset of summer strong high-level easterlies over Malaya being due to a developmental process rather than to a seasonal displacement over the territory of a pre-existing belt of strong easterlies lying to the north or south in the winter months.

TABLE I—STATISTICS OF MONTHLY MEAN WINDS AT 40,000 FEET AT 0300 G.M.T. OVER SONGKHLA (*B*) AND SINGAPORE (*A*)

			No. of obs.	V_S	V_N	V_E	$\overline{V}$		σ	α
					knots		°	kt.	kt.	
Jan.	1955	*B*	11	22·0	−12·5	12·4	135	18	17·3	80
	1951–55	*A*	65	23·3	− 7·2	20·2	110	21	13·1	92
Feb.	1954–55	*B*	19	19·5	−10·1	3·7	160	11	19·3	55
	1951–55	*A*	78	28·0	−10·7	23·3	115	26	17·2	92
March	1954–55	*B*	27	16·8	− 6·8	6·5	135	9	16·4	56
	1951–55	*A*	85	20·8	− 6·4	15·1	110	16·5	16·2	79
April	1954	*B*	16	16·5	− 3·6	7·0	115	8	17·0	48
	1951–55	*A*	79	19·4	− 2·0	16·3	110	16	14·1	85
May	1954–55	*B*	39	22·8	6·2	18·4	70	19	18·7	85
	1951–55	*A*	81	24·0	3·7	20·2	80	20·5	16·2	86
June	1953–55	*B*	31	32·2	9·8	27·2	80	29	20·2	90
	1951–55	*A*	81	35·9	10·5	32·3	70	34	15·6	95
July	1953–55	*B*	64	41·7	14·7	35·3	70	38	19·1	92
	1951–55	*A*	84	40·2	9·5	36·8	75	38	16·4	95
Aug.	1953–55	*B*	45	39·2	11·6	36·1	70	38	16·7	97
	1951–55	*A*	77	43·1	11·7	39·7	75	41	16·7	96
Sept.	1953–54	*B*	29	40·4	12·5	36·9	70	39	19·1	97
	1951–52, 54–55	*A*	71	39·8	11·4	36·2	70	38	18·0	95
Oct.	1953–54	*B*	39	27·8	3·8	24·1	80	24	18·3	88
	1951–52, 54–55	*A*	81	26·3	4·4	24·1	80	24	15·0	93
Nov.	1954–55	*B*	27	26·4	− 6·2	21·2	105	22	17·8	84
	1951–55	*A*	96	26·5	− 1·6	24·7	95	25	14·3	93
Dec.	1954–55	*B*	43	24·6	−11·8	13·4	130	18	16·4	67
	1951–55	*A*	78	23·5	− 7·0	20·1	110	21	14·5	90

The analysis.—For the purpose of calculating stretch vector correlation coefficients it was decided, in view of the few simultaneous observations at Singapore and Songhkla in certain months, to combine the observations into four seasons. Because of the apparent marked change in wind régime from April to May, it seemed inappropriate to combine data for these two months: consequently the periods February–April, May–July, August–October and November–January were chosen.

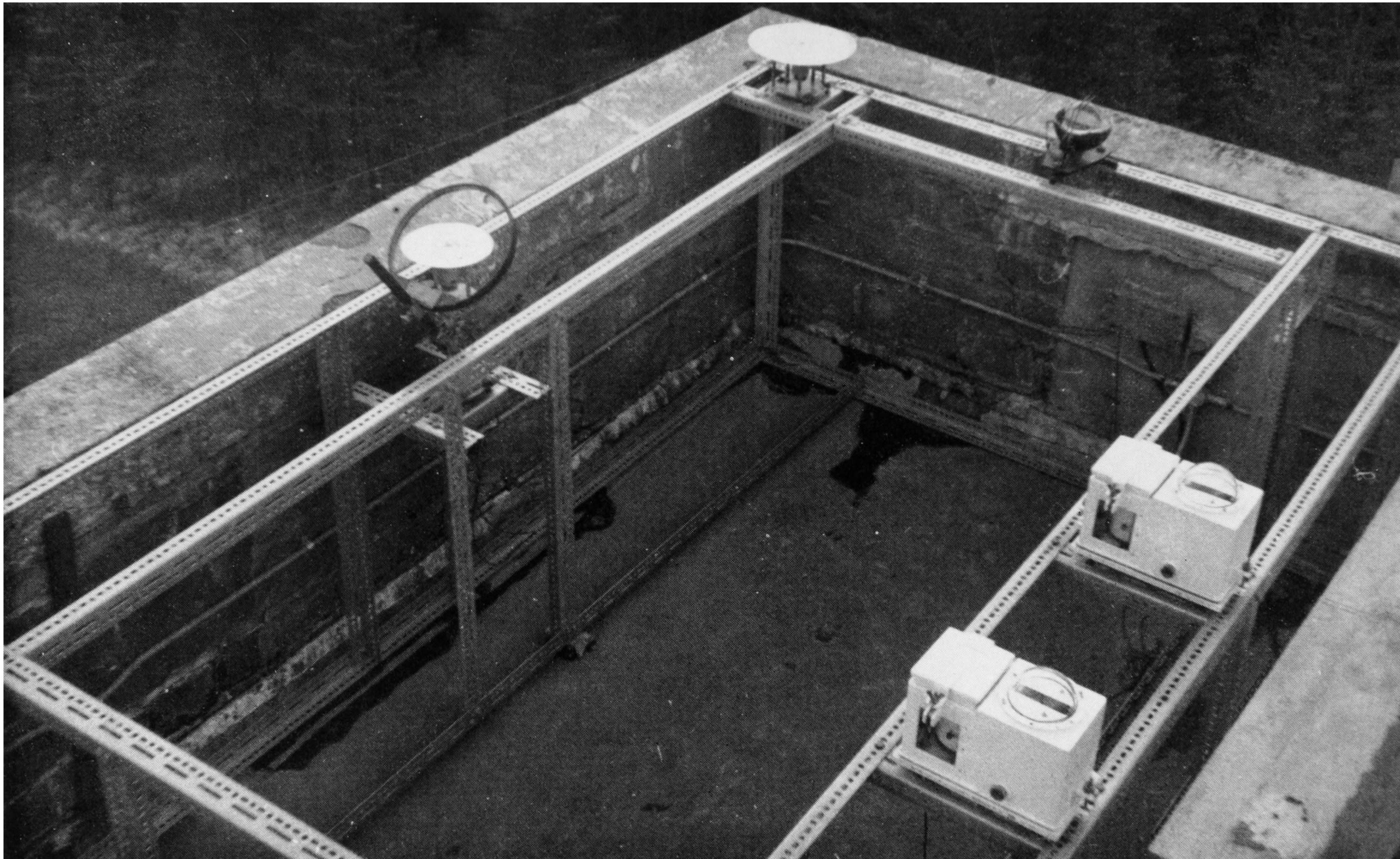
ESKDALEMUIR OBSERVATORY—GENERAL VIEW OF MAIN BUILDING

The workshop and laboratories are on the ground floor, offices and computing room on the first floor, and observational platform on the tower. In the foreground are the instrument screen, evaporation tank and cabinet containing apparatus for chemical sampling of air and rain-water. (see p. 129)

ESKDALEMUIR OBSERVATORY—FRONT VIEW OF MAIN BUILDING

Rayleigh House (Superintendent's quarters) is at extreme left.

(see p. 129)



OBSERVATIONAL PLATFORM ON TOWER

Total and diffuse radiation solarimeters, sunshine recorder and two bimetallic radiation recorders are shown.

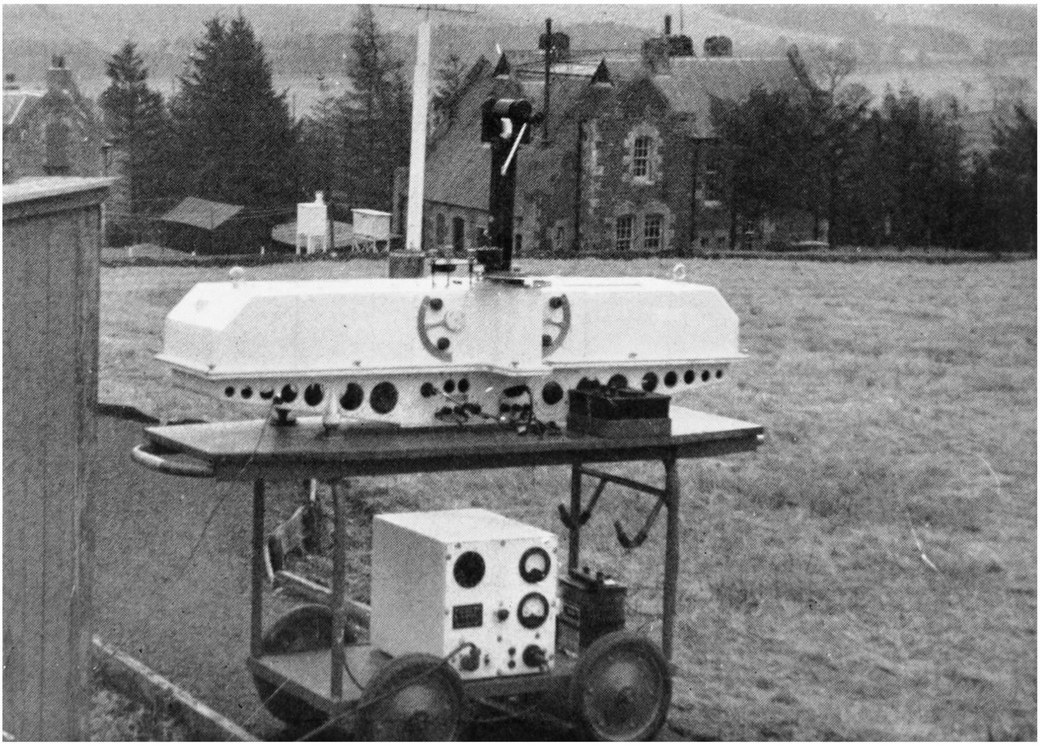
(see p. 129)

Crown copyright

RAIN-GAUGES

A view of some of the rain-gauges and experimental snow-gauges, with atmospheric electrical pit further back.

(see p. 129)



Crown copyright

OZONE SPECTROPHOTOMETER

Close-up of ozone spectrophotometer, showing Schuster House (hostel) in background.

(see p. 129)

For each period, vector means were derived from the north and east components of the individual observations, and standard vector deviations computed by finding the mean square of the modules of the vectors, subtracting the square of the module of the mean vector, and taking the square root of the resulting difference[10].

TABLE II—STATISTICS OF MONTHLY MEAN WINDS AT 50,000 FEET AT 0300 G.M.T. OVER SONGKHLA (*B*) AND SINGAPORE (*A*)

			No. of obs.	V_S	V_N	V_E	$\overline{V}$		σ	α
					knots		°	kt.	kt.	
Jan.	1955	*B*	11	19·9	− 0·5	16·6	90	17	15·8	83
	1951–55	*A*	64	39·7	− 6·0	36·3	100	37	22·1	93
Feb.	1954–55	*B*	19	25·0	− 3·9	6·1	125	7	29·8	28
	1951–55	*A*	75	38·6	− 7·7	28·7	100	30	33·9	77
March	1954–55	*B*	26	19·2	− 6·5	7·0	135	10	21·8	50
	1951–55	*A*	85	23·9	3·7	9·4	70	10	26·0	42
April	1954	*B*	10	19·6	− 3·6	7·5	115	8	21·2	42
	1951–55	*A*	78	23·9	2·9	19·5	80	20	19·6	82
May	1954–55	*B*	34	40·6	5·8	38·0	80	38	24·1	95
		A	78	30·9	9·0	26·0	70	27·5	19·3	89
June	1953–55	*B*	24	49·8	13·4	43·4	75	45	29·2	91
	1951–55	*A*	75	39·5	9·3	34·9	75	36	24·1	91
July	1953–55	*B*	59	61·0	17·4	55·4	75	58	30·4	95
	1951–55	*A*	76	43·3	8·6	39·9	80	41	23·2	94
Aug.	1953–55	*B*	46	52·3	10·8	49·3	80	51	27·3	97
	1951–55	*A*	64	52·3	12·0	49·4	75	51	22·0	97
Sept.	1953–54	*B*	25	57·0	3·7	52·7	85	53	28·6	93
	1951, 52, 54, 55	*A*	65	43·9	7·7	40·9	80	42	27·7	95
Oct.	1953–54	*B*	37	41·2	5·3	38·8	80	39	21·5	95
	1951, 52, 54, 55	*A*	77	38·9	5·1	36·7	80	37	19·2	95
Nov.	1954–55	*B*	24	33·2	−10·6	29·4	110	32	21·7	97
	1951–55	*A*	97	41·8	1·8	40·4	90	40	21·6	97
Dec.	1954–55	*B*	40	28·8	− 8·6	22·7	110	24	22·5	84
	1951–53, 55	*A*	70	33·6	− 4·1	29·5	100	30	23·8	89

For each pair of simultaneous observations at Singapore and Songhkla the vector difference was determined, and the root mean square, σ_x, computed for each season. Stretch vector correlation coefficients were then evaluated from the relation

$$r_x = \frac{\sigma_A{}^2 + \sigma_B{}^2 - \sigma_x{}^2}{2\sigma_A\sigma_B}$$

where suffix *A* refers to Singapore, and *B* to Songhkla.

The annual values and the 5 per cent confidence limits of r_x were obtained by applying Fisher's z' transformation in the way described by Brooks and Carruthers[10].

The results of this analysis are contained in Tables III and IV.

TABLE III—STATISTICS OF SEASONAL WINDS AT 40,000 FEET AT 0300 G.M.T. AT SONGKHLA (*B*) AND SINGAPORE (*A*)

		No. of obs.	$\overline{\mathbf{V}}$		σ	No. of pairs of obs.	Standard vector difference σ_x	Stretch vector correlation coefficient r_x	Confidence limits of r_x	Regression coefficient $\sigma_B / \sigma_A \, r_x$
			°	kt.	kt.					
Feb.–Apr.	*B*	62	140	9	17·7	61	18·8	0·40	0·17–0·59	0·42
	A	242	110	19	16·7					
May–July	*B*	134	70	31	20·8	116	21·3	0·40	0·24–0·54	0·47
	A	246	75	31	17·7					
Aug.–Oct.	*B*	113	75	33	19·4	67	19·2	0·48	0·27–0·65	0·51
	A	229	75	34	18·2					
Nov.–Jan.	*B*	81	120	19	17·7	52	19·6	0·27	0·00–0·51	0·33
	A	239	105	23	14·5					
Annual	...					296		0·40	0·30–0·49	

TABLE IV—STATISTICS OF SEASONAL WINDS AT 50,000 FEET AT 0300 G.M.T. AT SONGKHLA (*B*) AND SINGAPORE (*A*)

		No. of obs.	$\overline{\mathbf{V}}$		σ	No. of pairs of obs.	Standard vector difference σ_x	Stretch vector correlation coefficient r_x	Confidence limits of r_x	Regression coefficient $\sigma_B/\sigma_A \, r_x$
			°	kt.	kt.					
Feb.–Apr.	*B*	55	125	8	24·8	53	21·2	0·69	0·52–0·81	0·60
	A	238	90	19	28·6					
May–July	*B*	117	75	50	29·8	98	30·4	0·36	0·17–0·52	0·47
	A	229	75	35	23·0					
Aug.–Oct.	*B*	108	80	47	26·6	64	29·5	0·32	0·08–0·53	0·36
	A	206	80	43	23·8					
Nov.–Jan.	*B*	75	110	25	22·8	45	25·5	0·38	0·10–0·61	0·37
	A	231	93	36	23·2					
Annual	...					260		0·43	0·33–0·53	

Discussion.—When wind observations at 40,000 feet and 50,000 feet are lacking from Songhkla, the most probable high-level wind, $\mathbf{V}_B$, to be expected 400 nautical miles to the north-north-west of Singapore at the same time as an observed wind, $\mathbf{V}_A$, at Singapore may be determined by solving the vector regression equation

$$\mathbf{V}_B = \overline{\mathbf{V}}_B + {}^{\sigma_B}\!\Big/_{\sigma_A}\ r_x\,(\mathbf{V}_A - \overline{\mathbf{V}}_A)\ ,$$

appropriate values for $\overline{\mathbf{V}}_A$, $\overline{\mathbf{V}}_B$ and the regression coefficient

${}^{\sigma_B}\!\Big/_{\sigma_A}\ r_x$ being selected from Tables III and IV.

Because of the relatively small correlation between simultaneous high-level winds at Singapore and Songhkla in most seasons, adopting the appropriate vector mean for Songhkla as a prediction of the wind in that region will invoke a standard vector error, σ_B, generally only about 2 knots greater than the alternative of using the regression equation.

From the data in Tables III and IV there can be little doubt that Singapore and Songhkla are in the same high-level wind régime of steady strong east-north-easterlies from May to October: but within the general stream there are evidently variations on a horizontal scale slightly less than that of temperate latitude disturbances, such that simultaneously occurring winds at places separated by about 400 nautical miles lying normal to but in the same easterly high-level stream show only a rather small correlation.

Extension of the analysis.—It is desired to develop a statistical method of predicting the most probable vector wind at a place distant x nautical miles from Singapore at a time t hours after an observation of high-level wind over Singapore itself.

We have

$$\mathbf{V}_{A}\,,\mathbf{t} = \overline{\mathbf{V}}_A - r_t\,(\overline{\mathbf{V}}_A - \mathbf{V}_{A}\,,o)\ , \qquad \ldots\ldots (1)$$

where $\mathbf{V}_A$, $\mathbf{t}$ is the probable wind at Singapore t hours after an observed wind $\mathbf{V}_A$, o and r_t is the appropriate temporal vector correlation coefficient as evaluated earlier.[11]

Also,

$$\mathbf{V}_{B}\,,\mathbf{t} = \overline{\mathbf{V}}_B - {}^{\sigma_B}\!\Big/_{\sigma_A}\,.\,r_x\,(\overline{\mathbf{V}}_A - \mathbf{V}_{A}\,,\mathbf{t})\ , \qquad \ldots\ldots (2)$$

where $\mathbf{V}_B$, $\mathbf{t}$ is the probable wind at Songkhla t hours after an observed wind at Singapore, and r_x is the stretch vector correlation coefficient between simultaneously observed winds at, and at x nautical miles from, Singapore.

Eliminating $\mathbf{V}_A$, $\mathbf{t}$, we obtain

$$\mathbf{V}_{B}\,,\mathbf{t} = \overline{\mathbf{V}}_B - {}^{\sigma_B}\!\Big/_{\sigma_A}\,.\,r_x\,.\,r_t\,(\overline{\mathbf{V}}_A - \mathbf{V}_{A}\,,o). \qquad \ldots\ldots (3)$$

In most cases the vector mean wind and the standard vector deviation at place B will not be known exactly, but it is legitimate to assume that at least within about 400 miles of Singapore $\overline{\mathbf{V}}_B$ and σ_B will be approximately equal to $\overline{\mathbf{V}}_A$ and σ_A respectively.

With the assumption that $\overline{\mathbf{V}}_B \simeq \overline{\mathbf{V}}_A = \overline{\mathbf{V}}$ and $\sigma_A \simeq \sigma_B$, equation (3) becomes:—

$$\mathbf{V}_{B}\,,\mathbf{t} = \overline{\mathbf{V}}_B - r_x\,.\,r_t\,(\overline{\mathbf{V}}_A - \mathbf{V}_{A}\,,o)\ . \qquad \ldots\ldots (4)$$

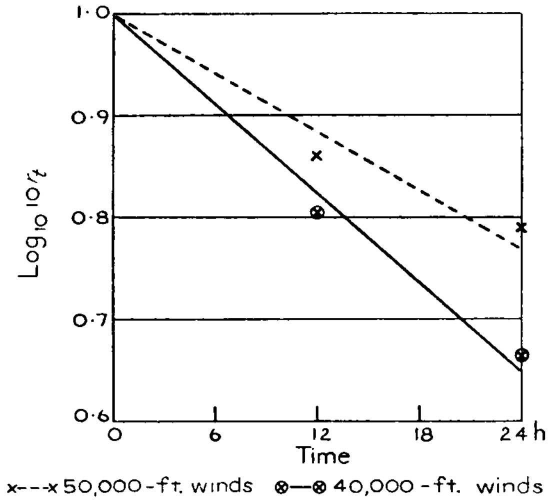


FIG. 1—PROBABLE VARIATION OF r_t WITH TIME OVER MALAYA

Mean values of r_t for t = 12 and 24 hours for Singapore have been found[11] to be:—

	r_{12}	r_{24}
50,000 ft.	0·73	0·615
40,000 ft.	0·64	0·46

Assuming, with Durst[5] a logarithmic variation of r_t with time, values of $\log_{10} 10 r_t$, and hence r_t, for various time intervals may be obtained from the graph at Figure 1 for winds at 40,000 feet and 50,000 feet over Singapore.

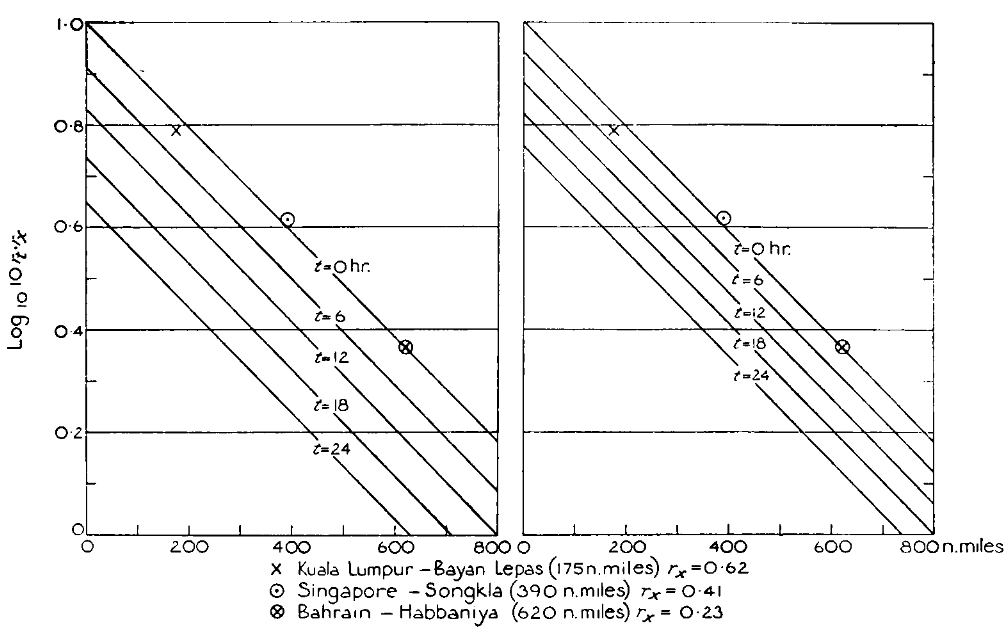


FIG. 2—PROBABLE VARIATION OF $r_t . r_x$ AT TWO HEIGHTS OVER MALAYA

Bearing in mind the confidence limits of the seasonal values found for r_x in Tables III and IV, a representative mean for r_x of 0·41 may be taken for a distance of 390 nautical miles at 40,000 feet and 50,000 feet. For 6,000-foot winds separated by 175 nautical miles in Malaya, Durst[5] quotes $r_x = 0 \cdot 62$, and also $r_x = 0 \cdot 23$ for 300-millibar winds between Bahrain and Habbaniya, 620 nautical miles apart. Converting these values to $\log_{10} 10 r_x$ and plotting against distance, the uppermost curves on each graph in Figure 2 have been drawn as the best-fitting straight line to represent the most probable variation of the stretch vector correlation coefficient with distance over Malaya.

From Figure 1 and the uppermost curves in Figure 2 the derived curves for the variation of $\log_{10} 10 r_x \,.\, r_t$ with distance over Malaya when $t = 6$, 12, 18 and 24 hours are plotted at Figure 2 for the 40,000-foot level, and for 50,000 feet. Using these graphs and the vector means in Tables I and II values of $r_x \,.\, r_t$ and $\bar{\mathbf{V}}$ may be obtained for solving regression equation (4) by the convenient graphical method set out earlier.[11]

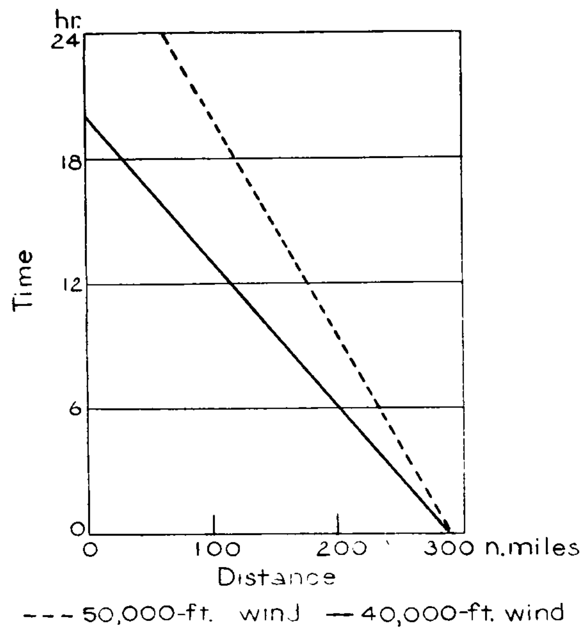


FIG. 3—LIMITING DISTANCES AND TIMES OUTSIDE WHICH THE VECTOR MEAN WIND IS LIKELY TO BE LESS IN ERROR AS A FORECAST THAN AN OBSERVATION AT SINGAPORE

Further discussion.—The standard vector error involved in using an observation $\mathbf{V}_{A,0}$ from Singapore as a forecast of the wind t hours later at a place distant x nautical miles is

$$E' = \sigma_A \sqrt{2(1 - r_x \,.\, r_t)} \;.$$

When $r_x \,.\, r_t$ is less than 0·5, this expression for E' is always greater than $E'' = \sigma_A$, the standard vector error of taking the vector mean as a forecast.

Now even for comparatively brief time intervals, Figure 2 shows that $r_x \,.\, r_t$ becomes less than 0·5, or $\log_{10} 10 r_x \,.\, r_t < 0 \cdot 7$, at quite short distances from Singapore. Thus the fairly common practice of using a stale radar wind observation for Singapore as a prediction of the wind at places 200–300 nautical miles or more away is not justified. Indeed, in all cases where the combination of staleness of observation and distance exceeds the limiting values in Figure 3, the unmodified observation is worse than useless as a forecast.

The standard vector error of predictions using equation (4) is

$$E''' = \sigma_A \sqrt{1 - (r_x \, . \, r_t)^2} \, ,$$

and this must always be less than $E'' = \sigma_A$ or the expression for E'. For example, from Figure 3, for $x = 150$ nautical miles and $t = 12$ hours, at 50,000 feet $r_x \, . \, r_t = 0{\cdot}53$, so that

$$E' = 0{\cdot}97\sigma_A$$
$$E'' = 1{\cdot}00\sigma_A$$
$$E''' = 0{\cdot}72\sigma_A \, .$$

Conclusions.—The high-level easterlies at Songkhla and Singapore show broadly similar features: their sudden onset in strength in May appears to be the result of development rather than advection.

In view of the lack of synoptic dynamical principles for wind forecasting and the observed rapid variation of wind with time and particularly distance in Malaya, an application of Durst's regression technique offers at present the best practical method for high-level wind prediction in the equatorial region of south-east Asia.

REFERENCES

1. WATTS, I. E. M.; Equatorial Weather. London, 1955.
2. BJERKNES, V. et alii; Dynamic meteorology and hydrography, Part II. Kinematics, Washington D.C., 1911.
3. SANDSTRÖM, J. W.; Über die Bewegung der Flüssigkeiten. *Ann. Hydrogr., Berlin,* **37,** 1909, p. 242.
4. PALMER, C. E.; Tropical meteorology. *Quart. J. R. met. Soc., London,* **78,** 1952, p. 126.
5. DURST, C. S.; Variation of wind with time and distance. *Geophys. Mem., London,* **12,** No. 93, 1954.
6. JOHNSON, D. H.; The success achieved in forecasting upper winds by orthodox and statistical techniques. *Met. Res. Pap., London,* No. 953, 1955.
7. CLARKSON, L. S.; Analysis of winds at 40,000 ft. and 50,000 ft. over Singapore. *Met. Mag., London,* **85,** 1956, p. 1.
8. HAY, R. F. M.; Winds at high levels over Singapore (1950–52). *Met. Res. Pap., London,* No. 770, 1952.
9. GILCHRIST, A.; Winds between 300 and 100 mb. in the tropics and subtropics. *Met. Rep., London,* **2,** No. 16, 1955.
10. BROOKS, C. E. P. and CARRUTHERS, N.; Handbook of statistical methods in meteorology. London, 1953, pp. 196, 222.
11. CLARKSON, L. S.; Variation with time of winds at 40,000 ft. and 50,000 ft. over Singapore. *Met. Mag., London,* **85,** 1956, p. 99.

METEOROLOGICAL OFFICE DISCUSSION

Forecasting low cloud conditions at airfields

The subject of the Monday Discussion of January 20, 1958 was dealt with under the following four headings by the openers noted against each one:

Frontal cloud: C. P. Soppet, B.Sc.

Orographic cloud: H. F. Hollands.

North Sea stratus: D. W. Johnston, B.Sc.

Non-frontal cloud in south-westerly airstreams: T. Johnston.

A discussion was held on each part after the opener's statement.

Frontal cloud

Mr. C. P. Soppet opened with a discussion of low cloud associated with warm and cold fronts approaching south-west England from a westerly quarter. The problem is primarily the successful forecasting of movements of fronts requiring current charts carefully analysed, particularly in regard to minor discontinuities, and a sound appreciation of likely development over the forecast period.

The differences of technique between coastal and inland stations were illustrated by reference to St. Mawgan and Exeter respectively.

Warm frontal low cloud at St. Mawgan is adequately forecast by advecting the cloud sequence over the sea or the change in the condensation level at the speed of the front.[1, 2] The important feature is the "stepped" character of these changes. Corrections are made for trajectory and wind speed. Diurnal and seasonal effects are small and little or no improvement occurs at the passage of warm fronts.

At Exeter shelter effects are considerable and upstream cloud reports are of little value. Lapse and hydrolapse rates upstream are useful for short period forecasts. Shelter effects are much reduced with slow moving fronts. Diurnal and seasonal effects are considerable. In summer cloud base is not very low and there is usually a rapid clearance following a warm front passage. In winter when the ground is frozen, very low cloud develops abnormally far ahead of a warm front and clearance to the rear is slow.

Cold fronts usually move at the speed of the geostrophic component. Sansom's[3] study is a useful aid in deciding whether post cold-frontal cloud decrease is likely to be rapid or slow and for short term forecasting, reference is made to cloud sequences from Irish stations. Timing both onset and clearance on these considerations is normally satisfactory for Exeter but breaks down for the north coast of Cornwall where clearance is usually delayed and in practice found almost impossible to forecast. Convection off the sea was suggested as a cause for this. Not directly associated with cold fronts but of considerable significance is the rapid development of fog or low stratus in radiation conditions following cold front clearances.

Examples were given of how certain types of low-level circulations and general development affect the above general considerations. Meteorological characteristics of airfields, frequency tables and other statistical data are all useful but should be applied in the context that each situation is unique. While there are rules to cover some aspects of forecasting frontal low cloud, much depends on the experience and judgement of the forecaster especially in regard to early recognition of development and also to the correction to be applied on account of site of airfield etc.

In the subsequent discussion agreement was expressed by several speakers with Peterssen's views that a "stepped" rise of dew-point ahead of a warm front could be due to falling rain. Hodographs were mentioned as useful tools for frontal analysis.

Mr. Illsley, while agreeing with the significance of discontinuities in surface layers, could not agree to their being given the status of fronts and said that at the Central Forecasting Office, forecasters could be concerned with "major" fronts only.

Several speakers, including Mr. Lamb, agreed with the difficulty of forecasting the clearance following a cold front but considered that orographic effects were the major cause of delay. In tracing weak fronts it was suggested that examination of wet bulb potential temperatures at 850 millibars would be useful.

Mr. Bradbury enquired how an increase in humidity over the English Channel was allowed for in a southerly airstream from France. In reply Mr. Soppet gave illustrations of how low-level circulations affect this issue. Another speaker suggested that the predominance of south to south-east winds at Exeter may be due to the deflection of south-westerlies by the Dartmoor mass.

Orographic cloud

The process of formation of orographic cloud is that of cooling by adiabatic expansion. For forecasting its occurrence it is necessary to know the moisture content of the air and the amount of lifting which will be given to it. The direction of the wind is clearly important, also. It is normal in determining the lifting condensation level to assume an adiabatic lapse rate of 5·4°F. per thousand feet, using the surface temperature and dew-point and assuming also that there is no entrainment. Precipitation on windward slopes gives a lowering of the cloud base due to the evaporation of water into the lower layers. To the lee of high ground the loss of moisture will be shown in the lowering of the dew-point and the raising of the condensation level.

Manchester shows some orographic effect in north-westerly winds, with larger amounts of cloud and a lower cloud base than places near the coast. It also has a somewhat greater frequency of showers in an unstable air mass and more prolonged precipitation, particularly of drizzle, in a stable air mass.

At Ronaldsway cloud may form over the low hills to the west and south-west. Over a very small range of wind direction this cloud may stream downwind over the airfield itself.

Over Anglesey low cloud in moist west or south-west airstreams will normally increase in amount and lower over the gently rising ground inland, particularly with a surface wind of 12 to 18 knots and the dew-point below the sea temperature. Here it is difficult to separate the orographic effect from the effect of turbulence. At Harlech, at 769 feet above mean sea level, separation of these effects seems possible at times when the calculated lifting by turbulence is insufficient by itself to reach the lifting condensation level. In such a case low cloud may be formed at Harlech by orographic uplift, but there will be none at Valley.

Frequency tables can be a useful forecasting aid. Low cloud frequency tables for Valley, for instance, show an almost complete absence of cloud below 1,000 feet in east or south-east winds, due to the shelter of the Welsh Mountains. Local knowledge, too, plays its part in the recognition of orographic effects and will be shown in area forecasts issued by most stations, but may not involve the use of any particular forecasting technique.

North Sea stratus

The next speaker, Mr. D. W. Johnston, dealt with the problem of North Sea stratus. He pointed out that stratus originating from the North Sea is a more serious menace to much of the country than stratus from other directions, because of lack of topographic protection. Owing to diurnal temperature variations there is a higher incidence of North Sea stratus by night and in the morning than in the afternoon, and owing to the seasonal lag of sea-surface temperature there is a relatively high incidence of stratus over the sea in spring and early summer.

Mr. Johnston then dealt with the manner of the onset of stratus in two simple types of situation. The first was that in which an air mass, initially cloud free, both over the land and over the sea, spreads inland from the North Sea during the cooling period. Cooling of the atmosphere originates at the ground, and is propagated upwards by turbulent diffusion through a layer, the depth of which depends largely on the wind speed. A lapse rate approximating to the dry adiabatic is maintained in the turbulent layer while it remains unsaturated, but if the temperature curve cuts the original dew-point curve, cloud will begin to form at this level, with saturated lapse rate thence to the top of the turbulent layer. Once the cloud has formed it quickly becomes a complete cover, and further cooling is at a much reduced rate causing a slow deterioration.

Since the stratus develops towards the top of the turbulent layer, its height should be closely related to the wind speed, so that, with a certain wind speed it should be possible to say that the stratus, if in fact it does occur, will develop at a certain corresponding height. An approximate forecasting rule embodying this idea is as follows:

Wind at anemometer level 10 knots: height of stratus 800 feet, and so on in proportion.

The temperature at which stratus should form can be found by drawing on the appropriate tephigram the dry adiabatic from the dew-point curve at the height of stratus formation down to the surface isobar. Assessments as to whether the temperature will fall to this value, and if so at what time, can be made using cooling curves or night minimum prediction formulae.

It was emphasized that the process outlined above is one of local development, and that cloud forming in this way does not spread downwind in any regular fashion, but can develop over large areas almost simultaneously.

The next type of situation to be dealt with was one with winds blowing inland off the North Sea and with clear skies over the land initially, but with stratus over the sea. The forecaster's first requirement for dealing with such a situation is a complete picture of cloud conditions over the North Sea. The information available obtained from light vessels, trawlers and various aircraft operating in the area, always falls very far short of this requirement. In the absence of all information regarding cloud conditions over the North Sea near our coasts, the forecaster has to make an estimate based on the characteristics of the air mass when it left the continental coast, the duration of its travel over the sea, and the sea-surface temperature. The important case is a spring or early summer situation with a warm air mass flowing over a relatively cold sea surface. In such a case with slow travel of the air mass, cloud could develop down to very low levels approaching or even, in extreme cases, reaching the sea surface. The reasons for this are:

(i) The turbulent layer is much shallower than with air flowing at the same speed over the land.

(ii) The process of development goes on continuously, day and night, whereas over the land it is essentially a nocturnal process.

(iii) Water vapour is being continuously fed upwards from the sea surface.

Frost has worked out formulae for the modification of a cold air mass flowing over a relatively warm sea surface, but this is not very relevant to the stratus problem, and the forecaster, in an extreme case of lack of actual observations, would have to rely on his judgment and experience.

A layer of low cloud over the sea will spread bodily inland at a time when the inland temperature has fallen to a favourable level. A good guide to this temperature value is the sea-surface temperature, and the time at which it will occur can be estimated from cooling curves. The cloud as it spreads in will remain at much the same height as over the sea—this often being at a much lower level than for cloud developing over the land by nocturnal cooling plus turbulence. A complication can arise with a zone ahead of the North Sea stratus proper where stratus forms by local development, as in the first case dealt with, and at a somewhat higher level than the North Sea stratus. This complication has the effect of making the cloud appear to spread in an erratic fashion. Stations on or near the coast can experience in this type of situation a deterioration quite early—possibly in the latter part of the afternoon.

Mr. Johnston finally dealt with the important, though less vital, problem of the dissipation of North Sea stratus. Briefly, the suggested method of forecasting dispersal consists of determining from the tephigram the surface temperature which corresponds with the potential temperature at the top of the cloud, which may be known from aircraft reports or can be deduced from a tephigram. The rate of rise of surface temperature has to be estimated by the forecaster, taking into account the season and the thickness of the cloud layer. The cloud clearance at coastal stations is somewhat delayed owing to the continuous replenishment of cool air at the relevant heights, but a clearance normally spreads from inland backwards against the wind, and often crosses the coast and a strip of coastal water, although the low stratus persists well out to sea.

In the discussion which followed, *Mr. Fox-Holmes* drew attention to the importance of keeping track of weak inactive fronts over the North Sea, as these frequently had a belt of very low stratus associated with them.

Mr. Alexander made the point that, for the Cambridge area, the fen country was an important feature which could often be treated as an effective water surface, so increasing the stratus threat.

Mr. Hastings said that he had found that Frost's formula for the modification of an air mass crossing the North Sea worked quite well for a warm air mass flowing over a relatively cold sea surface, as well as for the reverse. As regards the warming of the surface air under a stratus sheet, he was of the opinion that the actual amount of water in the cloud sheet was of importance, as well as its thickness.

Non-frontal cloud in south-westerly airstreams

Mr. T. Johnston opened the aspect of non-frontal low cloud in a south-westerly airstream by discussing the forecasting problem for a single airfield rather than an area, the airfield being Prestwick. Because the aircraft using the field normally plan their flight some 15 to 18 hours ahead the forecaster must endeavour to give fairly long-range notice of the onset of poor cloud conditions.

There are two main sources of information about low cloud at Prestwick. One is a statistical survey of the incidence of fog and low cloud by Mr. N. E. Davis.[4] The other is a local, unpublished project into the problem of forecasting low cloud at the station. This local work being the basis of this opening statement, the speaker indicated the scope of the work. The approach was primarily synoptic on the grounds that practical forecasting usually begins with a prebaratic of some kind. All occasions of cloud at 600 feet or below were studied. Charts were examined and details of temperature, winds and clouds were recorded. The investigation covered two summers and one winter, from May 1949 to October 1950. The required conditions occurred on only 4 per cent of occasions. This figure was almost constant season by season and compares well with the longer-term average. Over half of the low cloud periods were directly associated with fronts and the number of air-mass cases was therefore limited, but certain features were common to both frontal and air-mass types.

After illustrating the topography of the area and pointing out the considerable shelter afforded by hills to the south and south-south-west of the field the speaker showed a diagram illustrating the strong predilection of very low stratus for a very limited range of surface wind direction. Well over half occurred in the range 230–250 degrees, with a rapid fall off to negligible figures on either side. The annual distribution of wind direction, shown on the diagram also, does not show a comparable extreme and the prevailing direction in fact lies at 180 degrees.

A further diagram showed the relation of the 230–250 degree range and the associated gradient range of 250–280 degrees in relation to the geography of the area. The former gives the longest possible unbroken sea track being reduced to one third or less by a 10-degree change outside it. The latter is the least broken flow from the Atlantic and again a small change outside the limits introduces large land masses upwind. Sea and air temperature comparison did not give very firm results, but it seemed that for cloud at 600 feet or lower the air mass should arrive in the Firth of Clyde with a dew-point not less than the sea temperature measured there daily.

The problem of cloud height was then noted and the great fluctuations over short periods of time discussed, and illustrated, to show the complexities and apparent intractibility of this aspect of the problem.

The speaker then summarised the tentative rules for the forecasting of cloud at 600 feet or below which were formulated as a result of the study. The most important first step is to make a prebaratic for the local area of an accuracy sufficient to give the gradient in the range 260–280 degrees. For an estimate of expected dew-point recourse must then be made to upwind ship reports, making adjustments for observed changes as the air moves northwards, inspection suggesting a fall of 3 to 4 degrees from position Juliett to the Firth of Clyde area. If the prebaratic can be made sufficiently accurate to about 25°W. it should be possible to differentiate between air with a source in the Biscay–Finisterre region which is usually too dry for low cloud and a source in mid-Atlantic which is much more likely to be sufficiently moist. For reasons mentioned earlier no firm aid could be found for forecasting cloud base beyond "600 feet or below" at any worthwhile range of time.

After pointing out that the remarks had been confined to a moving air mass and not to stagnant moist conditions which raised new issues and complications, the speaker then illustrated a recent random case, with the relevant charts and a description of events which conformed, in the important aspects, to the findings previously discussed.

REFERENCES

1. HINKEL, C. H. and SAUNDERS, W. E.; Speed of movement of warm fronts on the Atlantic. *Met. Mag., London,* **84,** 1955, p. 241
2. SUMNER, E. J.; A study of blocking in the Atlantic–European sector of the northern hemisphere. *Quart. J. R. met. Soc., London,* **80,** 1954, p. 402.
3. SANSOM, H. W.; A study of cold fronts over the British Isles. *Quart. J. R. met. Soc., London,* **77,** 1951, p. 96.
4. DAVIS, N. E.; Fog and low stratus at Prestwick Airport, with notes on the diurnal variation of surface wind and temperature. *Met. Mag., London,* **81,** 1952, p. 231.

NOTES AND NEWS

The movement of precipitation belts as observed by radar

Mr. W. G. Harper has pointed out that the maximum correlation between the movement of precipitation belts and radar winds is, in our January number (page 19), incorrectly stated to be at 700 metres. This should, of course, be 700 millibars. It should also be made clear that the correlations are with the components of the winds across the belts.

METEOROLOGICAL OFFICE NEWS

Retirement.—Mr. W. Andrews, Senior Experimental Officer, retired on February 20, 1958. He joined the Office as a Technical Assistant in January 1920 after service in the Royal Flying Corps and Royal Air Force from 1917 until the end of 1919. Apart from a period between 1934 and 1937 in the British Climatology Division at Headquarters, his 38 years' service has been spent at aviation outstations, including a tour of duty overseas. From 1945 until his retirement he served at Mildenhall.

Academic successes.—Information has reached us that the following members of the staff have been successful in recent examinations. We offer them our congratulations.

General Certificate of Education (Advanced Level): C. Alderson, P. N. Brown, Miss P. J. Carter, B. A. Cole, L. Fletcher, M. F. Gaskin, R. P. Healey, Miss A. S. Hill, N. Holdsworth, R. P. Johnson, M. J. Kerley, M. J. Llewelyn, Miss C. Lynch, K. Oram, B. N. Parker, Miss J. L. Platt and Miss B. F. Smith.

WEATHER OF JANUARY 1958

Northern Hemisphere

Both the Icelandic low and the Azores anticyclone were east of their normal positions; the former was less deep and the latter weaker than normal. An area of low pressure extended from the eastern United States seaboard to the Azores, interrupting the subtropical high pressure belt normally extending across the Atlantic in these latitudes. The Siberian anticyclone was only slightly north of its normal position, but the Aleutian low was displaced well to the east and was markedly deeper than normal.

An area of positive pressure anomalies covered Greenland, north-eastern Canada and the north-western Atlantic, the anomalies reaching +8 millibars in the Davis Strait. Asiatic Russia had pressure anomalies up to +8 millibars, and +7 millibars was reported in extreme north-east Asia. Pressure was also above normal in the northern Sahara, but in the North Pacific there were large negative anomalies reaching −18 millibars near 50°N. 150°W. Pressure was also below normal in the Atlantic, an anomaly of −10 millibars being recorded at Bermuda.

In the North Pacific exceptionally strong surface air flow occurred on the north-west side of the Aleutian low. By contrast, westerly surface air flow which is normally present in the Atlantic at about 50°N. in January, was substantially reduced.

Over the whole of North America, as far south as 35°N., temperatures were above normal. East of the Rocky Mountains anomalies reached +9°C., and +5°C. was reported from the Canadian Arctic. Elsewhere, temperatures were near normal; anomalies did not exceed 3°C. anywhere over Europe, the Mediterranean, North Africa and the North Atlantic.

Rainfall was above average over most of western and central Europe, northern Siberia, the coastal regions of North America, and over north-west Africa. Elsewhere in Europe and in central North America there was below average rainfall, and in the extreme north-east of Asia the month was unusually dry.

WEATHER OF FEBRUARY 1958

Great Britain and Northern Ireland

In the British Isles the predominantly mild and very wet weather of February was broken by short, unusually cold brighter spells.

During the first four days of the month pressure was high to the south of England and weather over the country was cloudy and mild with occasional rain, mainly in the north and west. On the 5th cold north-westerly winds spread over the whole country behind a vigorous depression over the North Sea, bringing a general fall of temperature of 10–15°F. and snow showers to many places. Temperature, which had risen to 56°F. at Aberdeen on the 4th, was below 29°F. throughout the 6th and fell to 2°F. early the following morning. On the 7th–9th there were heavy falls of snow in many parts of the country; in Scotland most of the main roads were blocked and many villages isolated, while snow lay 6–12 inches deep in many parts of northern England. Snow was up to 2 inches deep locally in the south-east but milder air had spread into south-west England, and on the 7th, temperatures in Cornwall rose to 50°F. in sharp contrast to the near freezing temperatures over the rest of the country.

From 8th–10th the cold air retreated slowly northward, and the belt of rain and snow which marked its departure was followed on the 10th and 11th by general and locally heavy rain associated with a deep depression off Ireland. Wind reached gale force in western districts and there was a rapid rise in temperature, especially in the south; 56°F. was reached at Herne Bay on the 10th. Weather remained mild generally until the 16th. In some southern districts temperature did not fall below 50°F. day or night for three days and nights and on the 14th, the warmest day of the month, reached 59°F. at many places in southern England and 60°F. at Cambridge. On the 16th and 17th northerly winds brought a sharp fall in temperature, but this was only a temporary interruption, and for the next few days there was slight rain in many places with temperatures near normal.

On the 24th a vigorous depression from the Atlantic was situated off south-west Ireland and a well marked frontal belt extended from east to west across the British Isles; snow was falling in the cold easterly winds to the north of the front. The depression moved eastwards across southern England during the night and winds reached gale force locally; gusts of 60 knots were recorded at Scilly and Felixstowe. Cold weather spread to the whole country in the rear of the depression, and temperature was near or below freezing everywhere throughout the following day, with widespread snow. Snow lay 6–12 inches deep in many parts of Scotland, northern England and the Midlands from 25th–27th and was 8 inches deep locally in Kent. On the 27th an anticyclone developed to the west of Ireland and milder air from the Atlantic spread over the country from the north-west, bringing widespread rain and drizzle and a general rise in temperature.

Mean temperature over the month was above average in the southern half of the country—about 2°F. above in the south-west of England—and below average further north—about 3°F. below normal in north-east Scotland. It was the wettest February over England and Wales since 1951 with 176 per cent of the average amount of precipitation. Twice the average was exceeded over the west coast of Wales, the Mersey Estuary, an area extending from mid-Wales to Warwickshire and over most eastern coastal districts. New records for February rainfall were established at Valley, Dishforth and Dyce.

The land was very wet in most places and outside work, including the planting of early potatoes, was held up. Winter spraying was rather behind schedule but glass-house work was going well. Cattle have, on the whole, come through the winter in good condition. Most districts seem to expect a good cropping season.

WEATHER OF MARCH 1958

The general character of the weather is shown by the following provisional figures :—

	AIR TEMPERATURE			RAINFALL		SUNSHINE
	Highest	Lowest	Difference from average daily mean†	Percentage of average*	No. of days difference from average*	Percentage of average†
	°F.	°F.	°F.	%		%
England and Wales ...	61	5	−4·0	83	−1	96
Scotland	59	−9	−4·4	89	−3	99
Northern Ireland ...	60	20	−4·0	63	0	88

*1916–1950 †1921–1950

RAINFALL OF MARCH 1958

Great Britain and Northern Ireland

County	Station	In.	*Per cent of Av.	County	Station	In.	*Per cent of Av.
London	Camden Square ...	1·19	75	*Carm.*	Pontcrynfe	1·04	28
Kent	Dover	2·26	126	*Pemb.*	Maenclochog, Dolwen Br.	2·07	53
,,	Edenbridge, Falconhurst	1·49	74	*Radnor*	Llandrindod Wells ...	1·03	40
Sussex	Compton, Compton Ho.	1·83	74	*Mont.*	Lake Vyrnwy	1·77	45
,,	Worthing, Beach Ho. Pk.	1·79	104	*Mer.*	Blaenau Festiniog ...	2·91	43
Hants	St. Catherine's L'thouse	1·50	77	,,	Aberdovey	1·30	49
,,	Southampton, East Pk.	1·53	73	*Carn.*	Llandudno	·69	37
,,	South Farnborough ...	1·84	108	*Angl.*	Llanerchymedd ...	·85	34
Herts.	Harpenden, Rothamsted	1·79	107	*I. Man*	Douglas, Borough Cem.	1·74	60
Bucks.	Slough, Upton ...	1·36	86	*Wigtown*	Newtown Stewart ...	1·23	40
Oxford	Oxford, Radcliffe ...	1·36	81	*Dumf.*	Dumfries, Crichton R.I.	1·54	56
N'hants.	Wellingboro' Swanspool	1·69	109	,,	Eskdalemuir Obsy. ...	1·94	49
Essex	Southend W.W. ...	·90	68	*Roxb.*	Crailing...	1·23	73
Suffolk	Ipswich, Belstead Hall	·96	66	*Peebles*	Stobo Castle	1·38	57
,,	Lowestoft Sec. School	1·59	120	*Berwick*	Marchmont House ...	2·79	132
,,	Bury St. Ed., Westley H.	·76	47	*E. Loth.*	N. Berwick	...	...
Norfolk	Sandringham Ho. Gdns.	1·41	89	*Midl'n.*	Edinburgh, Blackf'd H.	1·97	123
Dorset	Creech Grange... ...	1·68	67	*Lanark*	Hamilton W.W., T'nhill	·96	40
,,	Beaminster, East St. ...	1·69	63	*Ayr*	Prestwick	·93	44
Devon	Teignmouth, Den Gdns.	2·08	83	,,	Glen Afton, Ayr. San ...	1·22	33
,,	Ilfracombe	·93	38	*Renfrew*	Greenock, Prospect Hill	1·33	33
,,	Princetown	3·66	62	*Bute*	Rothesay, Ardencraig...	1·01	30
Cornwall	Bude	1·09	51	*Argyll*	Morven, Drimnin ...	...	...
,,	Penzance	4·64	146	,,	Poltalloch	...	...
,,	St. Austell	4·09	117	,,	Inveraray Castle ...	1·87	34
,,	Scilly, St. Mary ...	4·34	172	,,	Islay, Eallabus ...	1·47	47
Somerset	Bath	1·27	63	,,	Tiree	·75	29
,,	Taunton	·80	41	*Kinross*	Lock Leven Sluice ...	2·72	121
Glos.	Cirencester	1·56	71	*Fife*	Leuchars Airfield ...	4·07	250
Salop	Church Stretton ...	1·53	65	*Perth*	Loch Dhu	2·37	44
,,	Shrewsbury, Monkmore	·92	57	,,	Crieff, Strathearn Hyd.	1·93	77
Worcs.	Worcester, Diglis Lock	1·36	86	,,	Pitlochry, Fincastle	1·24	57
Warwick	Birmingham, Edgbaston	1·43	76	*Angus*	Montrose Hospital ...	5·06	307
Leics.	Thornton Reservoir ...	1·50	81	*Aberd.*	Braemar	3·11	139
Lincs.	Cranwell Airfield ...	1·50	103	,,	Dyce, Craibstone ...	4·77	234
,,	Skegness, Marine Gdns.	1·79	137	,,	New Deer School House	4·01	181
Notts.	Mansfield, Carr Bank...	1·81	103	*Moray*	Gordon Castle	1·66	98
Derby	Buxton, Terrace Slopes	2·37	82	*Inverness*	Loch Ness, Garthbeg ...	2·12	85
Ches.	Bidston Observatory ...	1·34	82	,,	Fort William	2·25	44
,,	Manchester, Ringway...	1·76	99	,,	Skye, Duntulm... ...	2·63	81
Lancs.	Stonyhurst College ...	1·66	63	,,	Benbecula	1·80	72
,,	Squires Gate	1·32	70	R. & C.	Fearn, Geanies ...	·95	77
Yorks.	Wakefield, Clarence Pk.	1·73	101	,,	Inverbroom, Glackour...	2·67	64
,,	Hull, Pearson Park ...	1·99	134	,,	Loch Duich, Ratagan...	3·05	59
,,	Felixkirk, Mt. St. John...	3·00	172	,,	Achnashellach	3·36	60
,,	York Museum	2·01	141	*Suth.*	Stornoway	2·13	93
,,	Scarborough	2·90	191	*Caith.*	Lairg, Crask	4·22	114
,,	Middlesbrough... ...	2·27	155	,,	Wick Airfield	2·93	162
,,	Baldersdale, Hury Res.	2·68	115	*Shetland*	Lerwick Observatory ...	1·65	53
Nor'l'd	Newcastle, Leazes Pk....	3·21	193	*Ferm.*	Belleek	2·50	90
,,	Bellingham, High Green	2·37	108	*Armagh*	Armagh Observatory ...	1·37	70
,,	Lilburn Tower Gdns ...	3·70	173	*Down*	Seaforde	2·24	83
Cumb.	Geltsdale	·78	34	*Antrim*	Aldergrove Airfield ...	1·46	73
,,	Keswick, High Hill ...	1·25	36	,,	Ballymena, Harryville...	1·57	59
,,	Ravenglass, The Grove	1·68	62	*L'derry*	Garvagh, Moneydig ...	1·51	57
Mon.	A'gavenney, Plâs Derwen	1·75	58	,,	Londonderry, Creggan	1·89	68
Glam.	Cardiff, Penylan ...	1·01	40	*Tyrone*	Omagh, Edenfel ...	1·07	40

* 1916–1950

Printed in Great Britain under the authority of Her Majesty's Stationery Office
By Geo. Gibbons Ltd., Leicester

Wt-18. K16.