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The variation of surface temperature with altitude

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Summary

Altitudinal gradients of temperature at screen level are determined by the lapse of temperature in the free atmosphere as modified by the surface. The effects of the surface can be considered in three stages:

(i) The diurnal and seasonal modifications to the lapse rate in the free atmosphere; these depend upon latitude, climate, and state of surface.

(ii) The influence of topography, where it is important to distinguish between local and large-scale effects.

(iii) Changes in climate and state of surface following the ground.

Although explicit consideration of these three factors does not necessarily provide a practical means of estimating surface lapse rates, it does enable the wide variety of observed values to be understood and placed in perspective.

1. Introduction

Altitude is one of the main causes of temperature variations near the surface of the earth. The change in temperature associated with a movement of several hundred kilometres in the horizontal may be accomplished by a change of only a few hundred metres in the vertical. Relatively few meteorological observations are made in mountainous districts and it is often necessary to estimate temperatures at a given location from observations made at a different altitude. As a result, it is important that differences of temperature introduced by altitude be well understood.

The variation of temperature with height in the troposphere is relatively simple. An upper limit to a decrease with altitude of 9.8 °C per km is provided by the dry adiabatic lapse rate resulting from the mixing of air unsaturated with water vapour. The saturated adiabatic lapse rate represents the effect of the mixing of cloudy air and varies with temperature according to the amount of moisture available for latent heat release. Values range from about half the dry adiabatic rate at low levels in the tropics to very close to the dry adiabatic rate near the tropopause, but 6 °C per km is a typical value. Whenever subsidence occurs in the atmosphere, more stable conditions prevail with isothermal and inversion layers being produced.

The final result of these factors is that mean temperatures in the troposphere decrease with altitude in a fashion similar to the saturated adiabatic lapse rate. Exceptions occur near the presence of semi-permanent subsidence inversions, such as occur over the subtropical oceans. The influence of the surface of the earth also causes modifications to lapse rates in the free atmosphere to occur on diurnal and seasonal time scales. Nevertheless, a lapse rate of 6 °C per km has become accepted as standard, and is commonly used for reducing temperatures to sea level.

The lapse rate of temperature over the ground, that is the variation with altitude of temperature recorded in a screen 1 to 2 metres above the ground, is not the same as that in the free atmosphere. It also depends on the effects of the surface, and these vary with latitude, climate, state of surface, local and large-scale topography. This large number of variables, and the interactions between them, produces a very wide range of lapse rates over the surface. The variations within the United Kingdom are illustrated by Smith (1975), who finds that lapse rates reach a seasonal maximum in spring on the northern Pennines, in summer on the eastern slopes of the Welsh hills, and in autumn on their western slopes.

The variations of surface temperature in mountainous districts have been studied for over a century. Most of the investigations have been local and have taken place in the Alps, where the number of observations available is greater than elsewhere. The reports of these studies are mainly contained in the German literature, but the findings of many of them are reported by Geiger (1965). A very extensive review of the literature is provided by Barry (1981).

Much of the work on the variation of temperature with altitude has been empirical, and the identification of the physical factors responsible for the observed variations in lapse rates has been of secondary importance — in many papers the contributions of the relevant variables have not been recognized explicitly. To the casual reader, therefore, the wide range of lapse rates reported, and the often inadequate physical explanations offered, present a very confusing picture. The main aim of this paper is therefore to present a simple and convincing account of the main factors involved in the determination of surface lapse rates, with special emphasis on the effects of large-scale topography. A general procedure for estimating surface lapse rates is suggested, but this is unlikely to be of direct practical value; it is offered mainly to act as a framework for a general understanding of the subject.

2. Geographical and seasonal variations of lapse rate in the free atmosphere

The continents respond to changes in solar radiation much more readily than the oceans so that winter lapse rates over the land are less than over the sea and decrease with latitude in response to the decreasing elevation of the sun. Over the oceans, however, lapse rates increase with latitude owing to the advection of progressively colder air from the land. In summer, lapse rates over the land are greater than over the sea and again decrease with latitude, but over the ocean latitudinal variations are slight. These changes are illustrated in Fig. 1, which displays profiles of mean temperature determined from radiosonde ascents for three land and three ocean stations. Variations over the sea are small compared to those over the land, where it can be seen that diurnal modifications to the atmosphere commonly extend up to about 1 km, and seasonal effects to 2 km. At Ocean Weather Ship (OWS) 'N', at 31 °N, some evidence of the subtropical subsidence inversion is present between 850mb and 700mb. Geographical and seasonal variations in lapse rate between 850–700 mb and 700–500 mb have been mapped globally by Maejima (1977).

Land temperatures respond fairly rapidly to seasonal variations in solar radiation, but time is required to transmit these changes to the atmosphere and ocean. Thus while the seasonal variations in temperature of the sea surface and atmosphere (as represented by the 1000–500 mb thicknesses) are approximately in phase, they are about a month behind those occurring over the land. This has the effect of making the atmosphere over the land more unstable in the spring than in the autumn. The differences, however, are small compared with those between winter and summer.

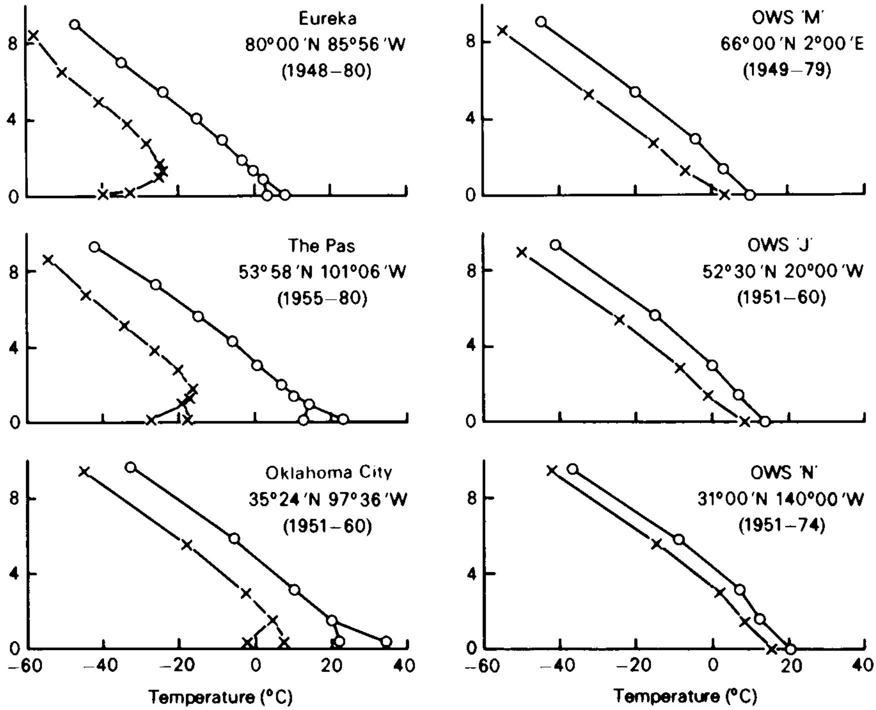

Figure 1. Profiles of mean temperature for stated periods. x—x January o—o July.

3. Forced ascent

The lapse rate following the surface of the earth is not the same as that in the free atmosphere; it also depends on the effects of the surface, and the most direct of these is that of forced ascent. When air is forced either up (or down) by the surface, its temperature changes either at the dry or saturated adiabatic rate, irrespective of that in the free atmosphere. The effect can be very important on individual occasions — föhn events, for instance, and when the melting of falling snow has produced an isothermal layer down to the surface. On these occasions, amounts of snow lying on the lowest ground may be small, but forced ascent will produce sub-zero temperatures on relatively high ground, leading to the accumulation of substantial amounts of snow or glaze at modest altitudes (see, for example, Pedgley 1969).

For mean temperatures, however, with which this paper is mainly concerned, forced ascent is not as important as appears at first sight. Surface adiabatic lapse rates caused by forced ascent will differ most from those in the free atmosphere when the latter are stable. Under these conditions, very little forced ascent is likely, either because winds are very light or airflow is occurring around, rather than over, the mountains. Substantial forced ascent is only likely when the lapse rates of temperature in the free atmosphere are close to adiabatic values, and hence when the modifying effects are small. Forced ascent and descent may, however, be responsible for modifying lapse rates by affecting the distribution of cloud and rain, but this is considered later under climatic effects.

4. Local and large-scale topography

Differences between lapse rates following the surface and those in the free atmosphere are mainly due to the effect of radiation at the surface, and this varies on a diurnal and seasonal time scale. The extent to which surface lapse rates are affected depends on the topography, and in this respect it is important to distinguish between the effects of local and large-scale topography. The term 'local topography' is used to describe the differences between hills and valleys etc. while the term 'large-scale topography' is used to refer to differences in altitude averaged over large areas, e.g. to distinguish between plateaux and plains.

Local topography is responsible for introducing differences in the lapse rate between maximum and minimum temperatures. A low-level station is usually in a valley, while a high-level station is commonly on a summit. The diurnal range in a valley will clearly be greater than that on a summit, and so the lapse of maximum temperatures must be greater than that of minima. Suppose, however, two sites of similar local topography are considered, one on a plain and the other on a plateau. The diurnal range of temperature at the two stations would be similar, and so the lapse rates of maximum and minimum temperatures would be the same.

The effects of local topography are not restricted to those affecting lapses of maximum and minimum temperature. The lapse of mean temperature in the diurnally modified atmosphere is often not the same as that in the seasonally modified atmosphere above it. In daytime in winter, the nocturnal inversion is not properly broken down and the diurnally modified atmosphere is more stable than the overlying layers. The extent of this modification will depend on the local topography and will be greater over a valley than over a plain. The phenomenon may be regarded as the development of a seasonal modification of the 'enclosed' atmospheres in valleys but, because it is a function of local topography, it is for convenience regarded as part of the 'diurnally' modified atmosphere. The seasonally modified atmosphere is intended to refer to the 'free' atmosphere above a level surface rather than the 'enclosed' atmosphere in a valley.

When considering altitudinal gradients of temperature measured along the ground, it is helpful to refer to three topographic models, and these are illustrated in Fig. 2. The first relates to local, and the second and third to large-scale, topography. The differences between the models are illustrated using a schematic temperature profile appropriate to a clear winter's night. Well above the surface a lapse of 6 °C per km is assumed; below this the seasonally and diurnally modified atmospheres are represented by isothermal and inversion layers respectively.

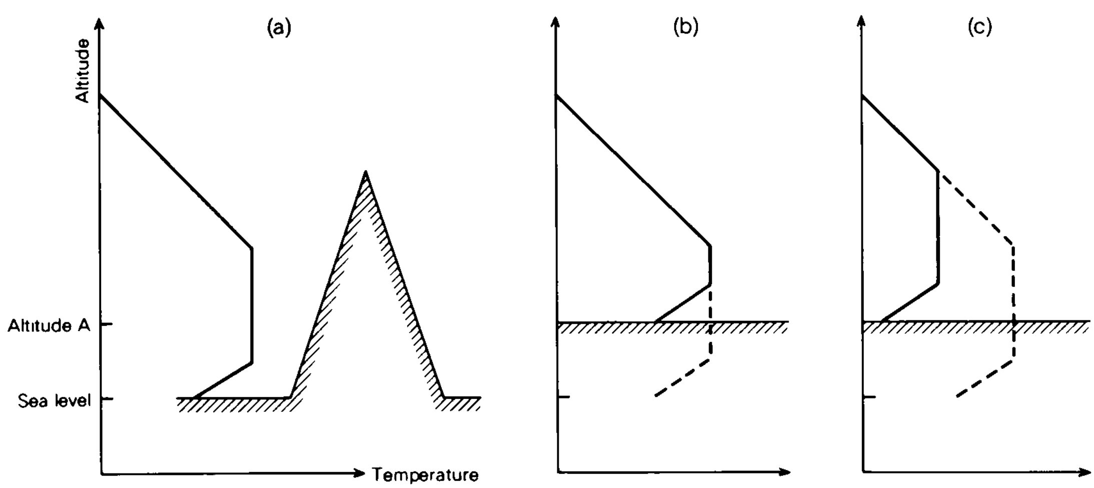

Figure 2. Schematic representation of temperature profiles associated with three models of topography — (a) isolated mountain, (b) limited plateau and (c) extensive plateau. ////////// land surface, ——— profile in free atmosphere above land surface, — — — profile if land surface was at sea level.

The first model is of a general lowland with isolated mountains projecting into the free atmosphere. The variations in temperature on the surface will then be similar to those observed in the diurnally and seasonally modified atmosphere. The second model is that of a small-scale plateau which projects into the seasonally modified atmosphere without extending it vertically. Lapse rates across the surface will then be similar to those in the seasonally (but not diurnally) modified free atmosphere. The third model is that of a plateau so extensive that the seasonally modified atmosphere extends as far above the plateau as above the lowlands. Differences in temperature between the plateau and those that would have been observed at sea level are then determined by the lapse in the unmodified atmosphere, i.e. around 6 °C per km. These last two models apply to gently sloping, as well as level, plateaux.

From a comparison of temperatures at sea level with those at an altitude A corresponding to the height of the plateau in models (b) and (c), Fig. 2 shows that for a winter's night:

(i) on an isolated mountain (a), temperature increases with altitude (the diurnally and seasonally modified lapse rates are relevant),

(ii) on a limited plateau (b), temperatures are independent of altitude (the seasonally modified lapse rate is relevant,

(iii) on an extensive plateau (c), temperatures decrease with altitude as the plateau is raised or lowered (the standard lapse rate of 6 °C per km is relevant).

A plateau so extensive as to build a seasonally modified atmosphere of the same depth as that which would occur over a comparable area of lowland (as in model (c)) probably does not exist on earth. The seasonally modified atmosphere, however, will always extend to greater altitudes over high ground than over lowland, so most areas of high ground will have characteristics intermediate between models (b) and (c).

A generalized model of the effects of large-scale and local topography is depicted in Fig. 3. A large

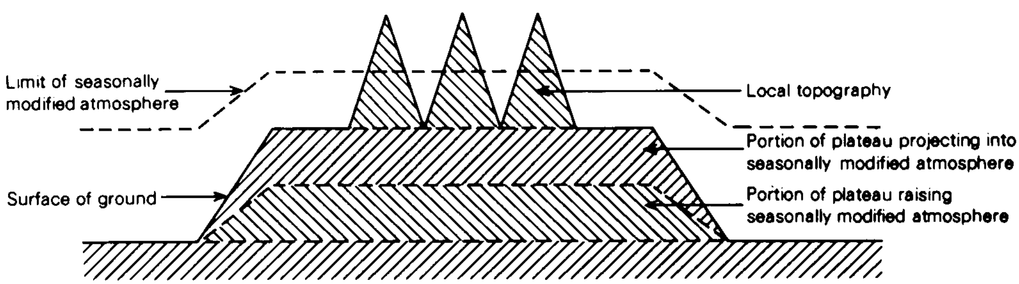

Figure 3. Composite model of topography for estimating surface temperature.

area of high ground will raise the height of the seasonally modified atmosphere, and to this extent may be regarded as belonging to the extensive plateau model of topography. Temperatures will decrease by around 6 °C per km up to the height by which the seasonally modified atmosphere has been raised. The remainder of the high ground may be regarded as projecting into the seasonally modified atmosphere, and lapse rates will be determined by these seasonally modified lapse rates in the free atmosphere. By averaging height over a certain area, a smoothly varying surface can be obtained and used to define the local topography. This is needed to introduce different lapse rates for maximum and minimum temperatures, and also to identify valleys which may contain seasonally modified 'enclosed' atmospheres.

The diurnal and seasonal modifications to lapse rates in the atmosphere are represented in Fig. 2 as being linear. In reality, of course, they will be non-linear, especially in winter, when lapse rates in the free atmosphere will change gradually from negative (i.e. inversion) near the surface to close to the standard value of 6 °C per km at higher levels. Thus on a gently sloping plateau conforming to model (b), for instance, increasing altitude would be associated with an increase in temperature at the lower levels which would be gradually retarded and eventually reversed at higher elevations.

5. Variation of radiation with altitude

The thinning of the atmosphere with increasing altitude has the effect of increasing both incoming and outgoing radiation. As a consequence, for topographically similar sites, diurnal range will increase slowly with altitude. As temperature decreases, however, outgoing radiation decreases according to Stefan's Law. This causes temperature to decrease slightly less rapidly with altitude over the land than in the atmosphere.

6. Coastal effects

Large changes in the diurnal and seasonal modifications to lapse rates in the free atmosphere occur in the vicinity of coasts. As air passes from sea to land or vice versa, changes in the lapse rate near the surface occur very rapidly (within a few minutes). With the passage of time these changes extend upward so that the diurnal modification is complete within a few hours. The seasonal modification is formed from the accumulated effect, over several days, of the diurnal changes.

Altitudinal gradients of temperature along the ground in coastal districts may be regarded as a combination of a horizontal coastal gradient and the vertical gradient that would have occurred inland. The horizontal coastal gradient will enhance the altitudinal fall in winter, but oppose it in summer. As a result, altitudinal gradients on coastal slopes will be similar to those that occur in the free atmosphere over the sea.

7. Climatic effects

Surface modifications to lapse rates in the free atmosphere are strongly dependent on the radiation received by the surface and this is a function of both climate and latitude. Thus differences in lapse rates of maximum and minimum temperatures induced by local topography will be much greater where the climate is clear and calm than where it is cloudy and windy. The strength of the stable layer which forms over continents in winter will depend on latitude and the frequency of clear, calm conditions. The extent to which these modifications affect lapse rates over the ground depends, as has been shown, on the topography. Thus for the extensive plateau model, differences in temperature from those that would have been observed at sea level depend only on the lapse rate in the unmodified free atmosphere, and are independent of climate. Note, however, that such a model requires the build-up of a large depth of seasonally modified atmosphere above the plateau and this is much more likely to occur if winds are light than if they are strong. Thus the extent to which a given plateau approaches this model may vary seasonally in accordance with variations in the strength of the wind.

Lapse rates following the ground will also depend on changes in climate, whether they be due to changes in the horizontal (e.g. distance from coast) or vertical (e.g. cloud cover). This will induce differences between lapse rates of maximum and minimum temperature, since the diurnal range in one type of climate will usually be different from that in another .Topography itself, of course, has a strong influence on climate. Thus lowlands on opposite sides of a mountain range often have differing degrees of continentality and exposure to given air masses. Isolated mountains and relatively small areas of high ground often have a cloudier, wetter, and windier climate than the adjacent lowlands. Extensive plateaux, on the other hand, often experience a dry, radiation-dominated climate. Mountain slopes on the windward edge of the plateaux often have a moist, cloudy climate due to the effects of forced ascent, but the air has lost most of its moisture by the time it arrives on the plateau.

Changes in the state of surface (vegetation and soil moisture) will also affect surface lapse rates through the proportion of radiation which is used to provide evaporation rather than surface heating, and there is an interaction with climate here. Thus, in summer, lapse rates in districts where cloud and rain increase with altitude will be greater than those where they decrease. The drier and sunnier location will not only receive more radiation, but the drier soil means that more of that energy will be available for surface heating.

8. Snow cover

The importance of the state of surface is also seen through the effects of snow cover which generally causes a reduction of temperature compared with that observed over exposed ground. Altitudinal gradients of surface temperature will be most affected when snow-free terrain at low levels is replaced by a snow cover at higher altitudes. The effects are likely to be greatest during the period when the snow is melting; some guide to the start of this period is given by the date of maximum snow depth. Using data from Geiger (1965), Table I implies that, in the Austrian Alps, this ranges from January near sea level to June above 3000 metres.

Table I. *Date of maximum snow depth in the Austrian Alps (Geiger 1965)*

Altitude (m)	200	600	1000	1400	1800	2200	2600	3000
Date	18 Jan	28 Jan	11 Feb	21 Feb	14 Mar	8 Apr	3 May	29 May

The evolution of seasonal variations in the depth of snow cover is illustrated by considering two stations whose mean temperatures differ by, say, 4 °C. During the winter, when temperatures at both stations remain below freezing, snow may be accumulating at the same rate at both stations. Suppose that the warmer station reaches its maximum snow depth in March. As spring advances, snow depth at this station decreases while that at the colder station still increases. By the time the snow depth at the colder station has reached its maximum, that at the warmer station may have completely disappeared.

Thus, during the early winter, altitudinal and latitudinal gradients of snow depth are relatively slight. As spring arrives, a steep gradient develops between those regions where the snow has disappeared and those where it has reached its maximum depth. As spring progresses, this boundary advances into regions of greater and greater snow depth, and the gradient of snow depth increases. By June, at high altitudes and high latitudes, snow depth will increase rapidly from zero to very high values. The snow in this transition zone will be melting rapidly and may be expected to exert a considerable influence on temperature. When the highest in a series of stations is located in this zone of melting snow, and a linear relation is fitted to temperature against height, a large value of the lapse rate is likely to result.

9. Case studies

9.1 *The Alps*

Some of the most extensive studies of mountain climate have taken place in the Alps, and some data given by Geiger (1965), and reproduced in Fig. 4, provide a good illustration of the effect of topography

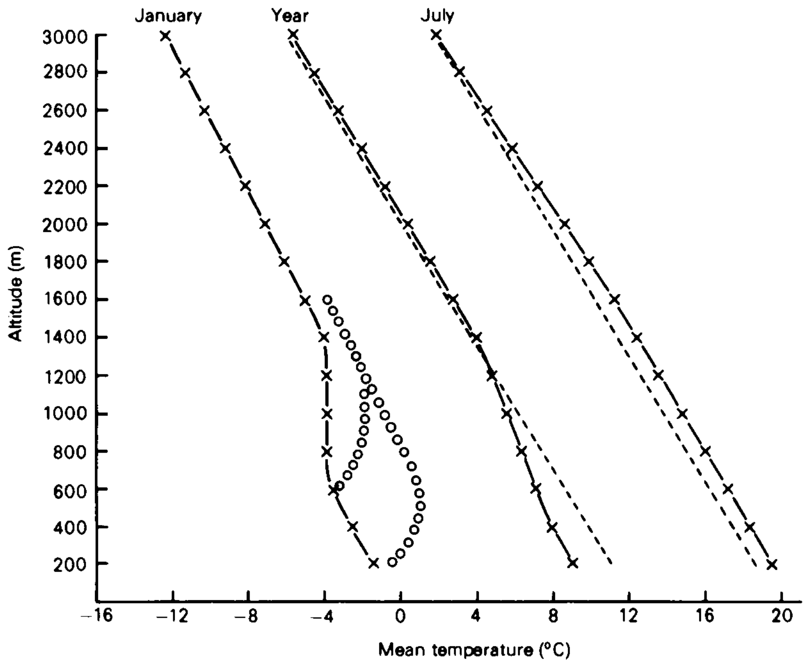

Figure 4. Variation of mean temperature with altitude in the Austrian Alps (Geiger 1965). x—x temperature, — — — lapse rate of 6 °C per km, ooo estimated temperature in atmosphere above surfaces at 200 m and 600 m.

on the variation of temperature with altitude. Fig. 4 displays the variation of mean temperature in January and July for various altitudes in the Austrian Alps. In summer, there is a reasonably linear relation between temperature and altitude with mean lapse rates close to 6 °C per km. In winter, however, an isothermal layer forms between 700 m and 1400 m. The average height of main Alpine valleys while they are still enclosed by mountains is represented by 700 m. Above 1400 m sites may be regarded as being on the open slopes of mountains projecting into the free atmosphere. Between 700 m and 1400 m sites may be regarded as being enclosed, with the atmosphere 'projecting' into the land. The observed variation of temperature with height is therefore very much a function of the average topographical exposure of a site at a particular altitude. The figures given for 1400 m, for example, would not apply to a valley at that level surrounded by high mountains.

Radiosonde ascents in the vicinity of the Alps show that, in July, temperatures in the free atmosphere are very similar to those displayed in Fig. 4. In January, however, ascents indicate that at the surface, 850 mb and 700 mb, temperatures are about 1 °C higher than those presented. The estimated profiles of mean January temperature in the atmosphere above a surface at 200 m outside the Alps, and above a valley at 600 m in the Alps, have been sketched. It is evident that in the region of the Alps, located towards the western margin of a continent, very little seasonal modification has taken place in the atmosphere, except that the lapse rate above 1400 m is rather less than 6 °C per km. Substantial modifications to mean temperatures extend only to about 500 m above the surface, and do not therefore reach beyond the range of diurnal modifications. The top of this layer, however, has been raised by about 500 m by the Alps, and so for the lowest 500 m temperatures decrease at approximately 6 °C per km in accordance with an extensive plateau model of topography. Thereafter the influence of local topography is dominant, and the effects of the 'diurnally' modified atmosphere, which includes a lowering of maximum as well as of minimum temperature in winter, represent the typical topographic exposure of a site at a given altitude.

It may be noted in passing that the depression of maximum temperature in Alpine valleys in winter is not solely a product of radiation; the melting of falling snow also makes a contribution. It is well known that when falling snow reaches the melting level, the energy extracted from the atmosphere by the melting can produce an isothermal layer close to 0 °C, and the snow gradually extends to lower and lower altitudes; the process is described by Lumb (1983). What is less widely appreciated is that such a mechanism will operate far more efficiently in light than in strong winds. While the melting of snow alone would produce an isothermal layer, the turbulent mixing caused by strong winds has the effect of producing an adiabatic lapse rate. Thus, in strong winds, the energy extracted from the atmosphere will be the same as in light winds, but the loss of energy will be redistributed vertically. With strong winds, there is also the possibility of warm advection at low levels, either from air that has been over the sea or drawn from outside the precipitation area. Low-level winds in the region of the Alps are often light, and the enclosed nature of Alpine valleys hinders large-scale mixing and encourages the development of an isothermal layer in the valley whenever falling snow reaches the melting level.

9.2 *Eastern slope of the Rockies*

One of the best examples of a gently sloping plateau occurs to the east of the Rocky Mountains in North America. Table II presents mean temperatures in January and July at two sets of stations, the

Table II. *Mean temperature (°C) over the period 1931–60 (with some shorter periods) at various stations on the sloping plateau to the east of the Rocky Mountains*

Station (50°N approx.)	Altitude (m)	Jan	Jul
Calgary	1079	- 10.0	16.7
Swift Current	816	- 13.1	18.9
Regina	574	- 16.5	19.0
Winnipeg	240	- 17.6	20.0
Station (40°N approx.)	Altitude (m)	Jan	Jul
Denver	1613	- 2.0	22.7
Sterling	1201	- 3.9	23.2
McCook	782	- 3.0	25.2
Concordia	448	- 2.6	26.8
Topeka	267	- 1.8	26.6

first, Calgary to Winnipeg, at around 50°N and the other, Denver to Topeka, at around 40°N. Although the precise figures for each station will be influenced by the local topography, the main factor is the large-scale topography. In winter, temperatures increase with altitude at around 9 °C per km at 50°N, while at 40°N they are approximately isothermal. Evidently, the second model of topography, in which the high ground projects into the seasonally modified atmosphere, is appropriate. In Fig. 1, the profile for The Pas, at 54 °N, shows that temperatures increase by around 4 °C in the lowest kilometre of the atmosphere. The seasonally modified atmosphere will be deeper at Winnipeg, near the continental interior, than at Calgary, and this accounts for the additional temperature difference between the two stations. At 40°N, the greater power of the sun prevents the formation of such large temperature inversions near the surface (see Oklahoma City in Fig. 1), and so isothermal conditions prevail. In summer, temperatures decrease with altitude at around 4 °C per km at 50°N and 3 °C per km at 40°N. Both these figures are less than the 'standard' rate of 6 °C per km. At 40°N, these differences are caused by changes in the climate, which becomes drier and sunnier as one moves west from Topeka towards Denver. The increased radiation and decreased evaporation at Denver permit higher surface temperatures than if the climate were the same as at Topeka, thereby decreasing the lapse rate.

9.3 *The Himalayas and Tibet*

The outstanding example of the effect of climate on lapse rates occurs in the region of the Himalayas. The Tibetan plateau is semi-arid all year, whereas the plains to the south pass dramatically from dry conditions in May to wet in July. Table III compares the lapses of maximum temperature between Leh and Lahore in the west and Lhasa and Patna in the east. The differences in temperature between the high- and low-level stations are almost halved between May and July as maximum temperatures on the lowland are depressed by cloud and rain.

Table III. *Lapse rate of maximum temperature (°C per km) in the Himalayan region (periods vary)*

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Leh–Lahore	6.6	6.6	6.4	6.6	7.3	6.4	3.9	3.7	4.6	6.1	6.1	6.2
Lhasa–Patna	4.4	4.6	5.7	6.0	5.2	3.4	2.6	2.6	3.1	4.1	4.3	4.1

Lapse rates of mean annual temperatures are considerably less than 6 °C per km, mainly due to the greater radiation and smaller evaporation at high levels in summer. These effects are opposed by the higher latitudes of the elevated stations and, especially in winter at Leh, to greater exposure to cold air masses from the north. The effect of the Tibetan plateau as a high-level heat source in summer is well known, and is discussed by Flohn (1953, 1968, 1974).

The variation of temperature following the ground in the Himalaya region behaves in a non-linear fashion. At low levels, maxima decrease rapidly as the plains are left for the hillsides, and cloud and rain increase. Then, as the Tibetan plateau is approached, temperatures rise as moisture and vegetation decrease. Maximum temperatures at Lhasa are actually higher than at Cherrapungi, 2500 m lower! These changes may be ascribed to horizontal gradients resulting from the change in climate, even though that change is largely caused by the topography in the first instance.

9.4 *The British Isles*

Good accounts of the variations of lapse rates in Britain are given by Manley (1970) and Harding (1978). Harding (1979) also gives a detailed discussion of the factors operating in the northern Pennines while Green and Harding (1980) analyse data in southern Norway to infer probable conditions in Scotland. With the exception of the Aberdeenshire plateau, the relief of Britain approaches that of the isolated mountain model of topography, and this forms the basis of the following description. Over the south-eastern half of England, the variation of solar radiation causes pronounced seasonal and diurnal variations in the lapse rate. Increased cloud and wind cause these variations to diminish towards the north-west. In winter, lapse rates over the seas to the west of Britain increase with latitude, and this, combined with the increased frequency of continental air masses towards the south-east, causes lapse rates over Britain to decrease from north-west to south-east. These differences are illustrated in Fig. 5 using data from Crawley and Stornoway, and OWS 'I' and 'J'. At Crawley, values exceeding 10 °C per km in summer are partly due to the lack of complete radiational screening of thermometers in a thermometer screen.

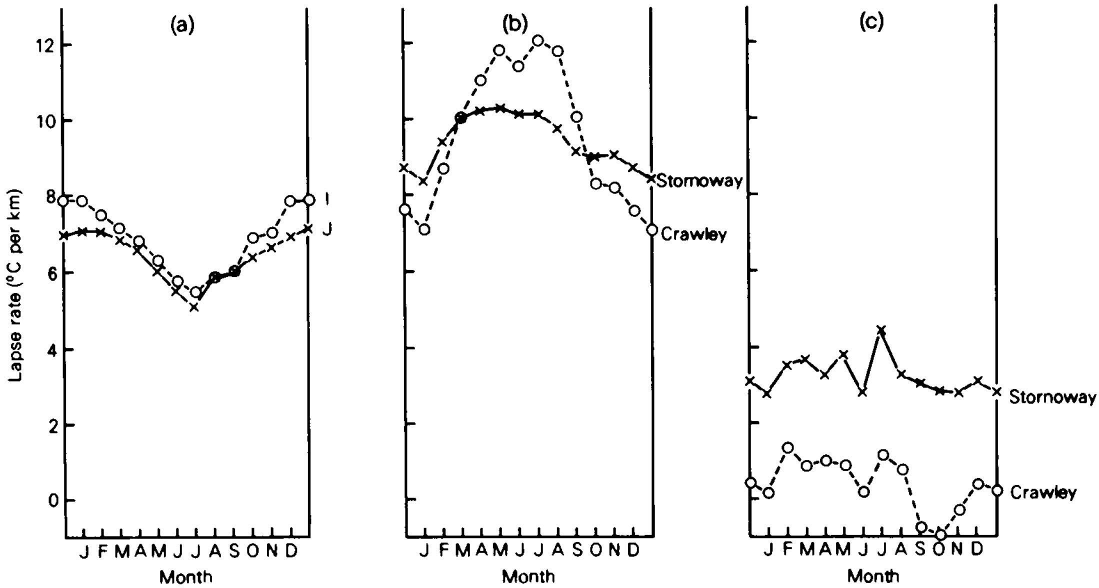

Figure 5. Mean lapse rates in the free atmosphere in the vicinity of the British Isles (1961–70). (a) Surface – 850 mb (b) Maximum temperature, surface – 900 mb. (c) Minimum temperature, surface – 900 mb.

According to Harding (1978) and Green and Harding (1980), the greatest lapse rates on Britain's northern hills are observed in spring. This is partly due to the melting of snow on the summits of these hills. The seasonal variation of lapse rates over the sea is opposite to that over the land, and this causes lapse rates measured along the coastal slopes of hills to reach a minimum in June and a maximum in December. The lapse of minimum temperature is non-linear, and increases gradually from near isothermal conditions in the lowest layers over the south-eastern half of England to close to 6 °C per km above about 1 km.

10. Conclusions

Lapse rates of mean temperature in the free atmosphere away from the influence of the surface are generally close to 6 °C per km, although there are exceptions near semi-permanent subsidence inversions such as occur in the zone of trade winds. Diurnal and seasonal modifications to this normal rate are caused by the effect of the surface, and the extent of these modifications depends on the latitude, climate, and state of surface.

Lapse rates of temperature over the ground can be estimated in three main stages:

(i) Take account of the location — the latitude, climate, and state of surface will determine the diurnally and seasonally modified lapse rates in the atmosphere.

(ii) Introduce the topography—part raises the seasonally modified atmosphere, part projects into the seasonally modified atmosphere, and part produces local effects.

(iii) Allow for changes in climate and state of surface following the ground — these are best regarded as being due to changes in the horizontal, even if they were largely caused by orography in the first place.

In more detail the recommended procedure for estimating altitudinal gradients along the ground is as follows:

(a) Obtain the seasonally modified lapse rates at various heights above the surface in the free atmosphere. These can be obtained from radiosonde ascents and, since they vary only on a large scale, can be mapped. For summer, and for the 850–700 mb and 700–500 mb, layers, this has already been done by Maejima (1977). Especially in winter, these lapse rates will be non-linear and will change as the surface is approached. In the zone of diurnal modification, however, lapse rates in the 'seasonally only' modified atmosphere should be regarded as linear extrapolations of those immediately above the zone.

(b) Estimate the height to which the seasonally modified atmosphere extends, and the amount by which it has been raised over areas of high ground. This may vary with season.

(c) Assume a lapse rate of 6 °C per km up to an elevation which equals the amount by which the seasonally modified atmosphere has been raised. Thereafter the lapse of temperature in the seasonally modified atmosphere should be used.

(d) Define a surface (e.g. by averaging height over a large area) which enables variations in height to be shared between a 'sloping plateau' and local topography. This will enable diurnal modifications to the free atmosphere, and seasonal modifications to 'enclosed' atmospheres to be taken into account.

(e) Allow for changes in climate and state of surface. By reducing temperatures to sea level using steps (a) to (d), these changes should be revealed as horizontal gradients, even though they were mainly caused by the orography.

The steps outlined above represent a generalized procedure which can only be implemented after a good deal of local study and empiricism, and are probably impossible to apply in any strictly objective and quantitative fashion. They are offered mainly to act as a framework for a general understanding of the subject and to enable diverse results from different parts of the world to be placed in perspective.

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The estimation of humidity parameters

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Summary

The calculation of various measures of humidity from wet- and dry-bulb temperatures is made difficult by the need to compute the saturation vapour pressure. Complex semi-empirical equations due to Goff and Gratch are available, but for many applications simpler and less accurate formulations are required. In meteorology, there is a need to calculate the dew-point from the wet-bulb temperature at synoptic observing stations, and this can be conveniently carried out on a programmable calculator. For this purpose an algebraically simple equation due to Magnus is shown to be most suitable, and procedures for its use in calculating humidity parameters at an observing station are described.

1. Introduction

The moisture content of the atmosphere is generally measured cheaply and conveniently by means of a wet- and dry-bulb psychrometer. Unfortunately, however, the calculation of alternative measures of humidity such as the dew-point (T_d), vapour pressure (e) and relative humidity is complex. The calculations are based on the Regnault equation,

$$e = e_s(T_w) - Ap(T - T_w),$$

where T = dry-bulb temperature,
T_w = wet-bulb temperature,
p = atmospheric pressure,
A = a ventilation coefficient, and
$e_s(T_w)$ = saturation vapour pressure at the wet-bulb temperature.†

The source of the computational problems is the saturation vapour pressure (e_s). This is related to temperature by an integration of the Clausius–Clapeyron equation, but this is made difficult by departures from the ideal gas law and the variation of latent heat with temperature. Consequently, semi-empirical formulae were derived by Goff and Gratch (1945) to enable e_s to be calculated to a high degree of accuracy. These equations have been subject to continual refinement and the latest versions are given by Wexler (1976, 1977). The Goff–Gratch equations apply, however, only to the saturation pressure of water vapour in the absence of other gases. When air is added to the water vapour e_s is increased, mainly owing to the extra forces of attraction between the molecules of air and water vapour. The magnitude of the effect at sea level pressures, however, is only about 0.5%. A complete account of humidity and moisture in the atmosphere is given by Wexler and Wildhack (1965), while a useful summary is provided by the World Meteorological Organization (1966).

The Goff–Gratch equations are relatively costly to evaluate on a regular basis and for many purposes they are unnecessarily accurate. Consequently many attempts have been made to devise empirical approximations in which simplicity is traded against accuracy. Some workers, e.g. Richards (1971), Lowe (1977) and Rasmussen (1978) have aimed for relatively high accuracy and developed high-order polynomials. The efforts of Richards and Lowe, together with some unpublished work by Hooper, are

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†Wherever it is important in this paper to distinguish between temperatures in the thermodynamic and Celsius scales, the former is denoted by T and the latter by t.

reviewed by Sargent (1980). Algebraically simpler but less accurate formulae were suggested by Tabata (1973), Revfeim and Jordan (1976), and Blackadar (1983), but the longest established of these simpler approximations is due to Magnus (1844). By incorporating an amendment due to Bogel (1979), Buck (1981) is able to recommend the use of the Magnus formula for a wide variety of uses and provides a choice of coefficients according to the temperature range of interest and the accuracy required. Buck also supplies a choice of equations for taking into account the 'enhancement factor' which represents the difference in saturation pressure between pure water vapour and moist air.

Many of the empirical approximations developed have assumed that the calculations will be performed on a large computer. In this context the simplicity of an equation has been interpreted in terms of its speed of execution. An IBM 3081 computer at the Meteorological Office, for instance, can evaluate a sixth order polynomial in about the same time as a single exponential expression. In operational meteorology, however, calculations have to be made at synoptic observing stations in order to convert from T_w (which is observed) to T_d (which is required by the World Meteorological Organization (WMO) synoptic codes). These calculations have generally been performed with the aid of either tables or slide-rule, but they may be made more conveniently on a microcomputer or programmable calculator. In this context algebraic simplicity is important as it minimizes storage requirements and enables an equation to be reversed in order to calculate temperature from saturation vapour pressure as well as vice versa.

For routine meteorological purposes, the accuracy required of e_s is that equivalent to an error of 0.1 °C in temperature. This is from 0.5% to 1%, and a number of algebraically simple expressions are capable of achieving this accuracy. A number of such formulae are compared, and the brevity, reversibility and accuracy of the Magnus formula are shown to make it ideal for use with a pocket calculator. It is also perfectly suitable for routine climatological applications on a large computer.

The precision to which T and T_d are reported in the WMO synoptic code messages was increased in 1982 from 1 °C to 0.1 °C. This increase in precision made it possible for climatological collecting centres to calculate the other humidity parameters direct from the messages, rather than through internal reporting of T_w. This procedure therefore offers the advantages of increased automation, but requires the recovery of T_w from T and T_d and this is not straightforward. The purpose of this paper is, therefore, threefold:

(i) to compare some of the algebraically simple formulae for calculating e_s,

(ii) to describe and recommend a procedure based on the Magnus formula for calculating humidity parameters from T and T_w,

(iii) to describe a method for recovering T_w from T and T_d.

The recommended procedures can then be implemented either on a large computer at a collecting centre, or on a programmable calculator at the observing site.

2. Integration of the Clausius–Clapeyron equation

The saturation vapour pressure is expressed as a function of absolute temperature by the Clausius-Clapeyron equation

$$\frac{1}{e_s}\,\frac{de_s}{dT} = \frac{EL}{RT^2}$$

where L = latent heat of vaporization of water = 2.50084×10^{-3} J kg^{-1} at 0 °C,

R = gas constant for dry air = 287.05 J kg^{-1},

and E = ratio of molecular weight of water vapour to that of dry air = 0.62198.

Difficulties in integrating this equation are caused by the fact that L is not constant, but varies with temperature. If this variation is ignored, and L is assumed to be constant then

$$e_s = \exp(21.4 - 5351\,T^{-1}) \quad \ldots \quad (1)$$

where e_s is given in millibars if T is expressed in kelvins. This is the equation used by Blackader (1983) to illustrate the calculation of humidity parameters on a home computer.

A better assumption is clearly to make L a linear function of T,

e.g. $$L = \{\, 2500.84 - 2.34\,(T - 273.15)\,\} \times 10^{-3}\ \mathrm{J\ kg^{-1}}$$

when $$e_s = \exp(55.17 - 6803\,T^{-1} - 5.07\,\ln T) \quad \ldots \quad (2)$$

The performance of equations (1) and (2), when assessed against the solution of the Goff-Gratch equations as given by WMO (1966), is illustrated in Fig. 1. In this diagram, the dotted lines indicate the

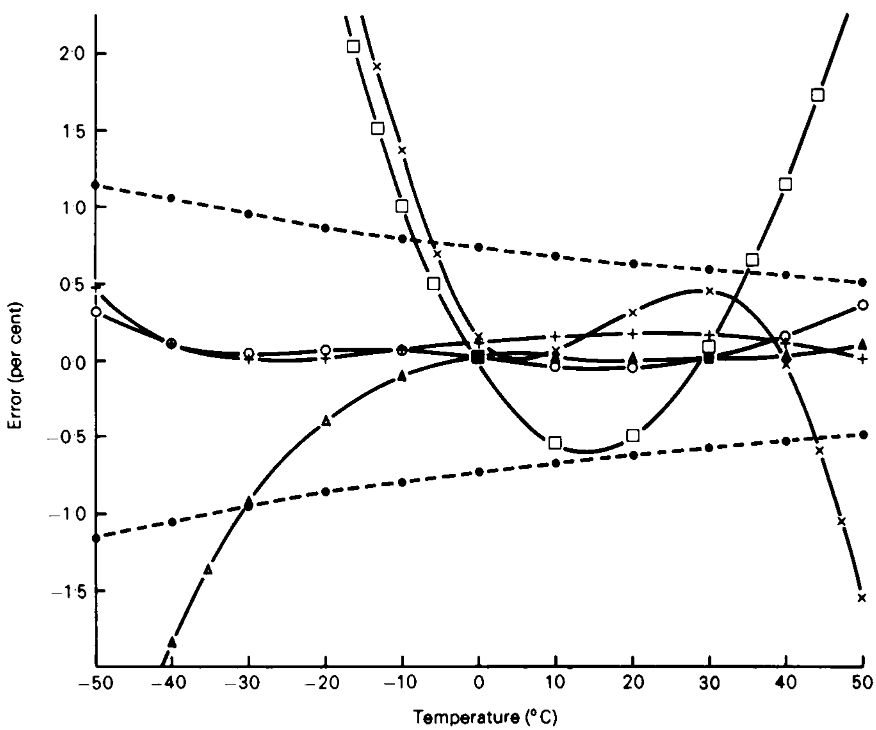

Figure 1. Error in saturation vapour pressure obtained from various formulae. ●- - -● error equivalent to 0.1 °C error in temperature □——□ Blackader, ×——× Revfeim and Jordan, △——△ Magnus, ○——○ Tabata, +——+ Clausius - Clapeyron with L a function of T.

error in e_s which is equivalent to an error in temperature of 0.1 °C. The assumption that L is constant is seen to produce errors within the required limits for temperatures between −10 °C and +35 °C. The assumption that L is a linear function of temperature, however, results in a considerable improvement and produces errors well within the required limits over the entire range of temperature examined.

3. Empirical expressions for calculating the saturation vapour pressure

The Magnus formula already referred to takes the form

$$e_s = 6.1070 \exp\left(\frac{at}{b+t}\right) \quad \ldots(3)$$

In this and all subsequent equations, vapour pressure is given in millibars if temperature is expressed in degrees Celsius. The most commonly quoted values of the coefficients for evaporation over water are $a = 17.3$ and $b = 237.3$. Minimum root-mean-square (r.m.s.) errors in millibars with respect to the Goff–Gratch values over the temperature range -40 °C to + 40 °C, however, are obtained by putting $a = 17.38$ and $b = 239.0$ for evaporation over water and $a = 22.44$, $b = 272.4$ for sublimation from ice. With these coefficients, Fig. 1 shows that the Magnus formula produces errors which are within the required range for all temperatures examined above -30 °C.

Revfeim and Jordan (1976) used the relation

$$e_s = \exp\{7.076 - 2.47\,(1.46 - 0.01t)^2\} \quad \ldots(4)$$

and Fig. 1 shows that this formulation achieves the required accuracy over a temperature range of -6 °C to +43 °C. Better results may be obtained by using a quadratic in the inverse of absolute temperature, T,

i.e. $$e_s = \exp(19.163 - 4063.2\,T^{-1} - 184089\,T^{-2}) \quad \ldots(5)$$

This form of equation was suggested by Tabata (1973), although the coefficients used are those which minimize r.m.s. errors (in mb) between ±40 °C. The accuracy of this equation is demonstrated in Fig. 1, where the errors produced can be seen to lie within the prescribed limits over the entire temperature range examined.

Of the five equations considered, Fig. 1 shows that those valid over the narrowest range of temperature are those suggested by Blackader (1983) and Revfeim and Jordan (1976). The widest range of applicability is attained by the equation due to Tabata (1973) and that obtained from an integration of the Clausius–Clapeyron equation with L a linear function of T. Unless very low temperatures are to be regularly dealt with, however, it is the Magnus equation which is recommended for use with pocket calculators. It has the advantage of being algebraically very simple and hence reversible, and, for temperatures between 0 °C and 40 °C, is the most accurate of all the equations examined.

The enhancement factor for moist air has not been used in the above calculations since the aim was to reproduce the solutions of the Goff–Gratch equations. It is also ignored in the following sections in order to maintain continuity with past and current Meteorological Office practice. Its omission does not lead to any errors in T_d or the relative humidity, while that in e_s is equivalent to an error in temperature measurement of just less than 0.1 °C. If it is wished to include the effect, however, this can be achieved simply by increasing e_s by 0.46%, i.e. by replacing the constant 6.1070 in the Magnus equation by 6.1351.

4. Recommended procedure for calculating humidity parameters from wet- and dry-bulb temperatures

The starting point is the calculation of e from the Regnault equation, which requires a knowledge not only of t and t_w but also of $e_s(t_w)$. This can be obtained from the Magnus equation by replacing t by t_w in equation (3) to give

$$e_s(t_w) = 6.1070 \exp\left(\frac{17.38t_w}{239.0+t_w}\right)$$

for $t_w \geqslant 0$ °C and

$$e_s(t_w) = 6.1070 \exp\left(\frac{22.44t_w}{272.4+t_w}\right)$$

for $t_w < 0$ °C.

Since $e = e_s(t_d)$, the Regnault equation may be written as

$$e_s(t_d) = e_s(t_w) - Ap(t - t_w).$$

For measurements in a screen of the Stevenson type the ventilation coefficient A is 0.000799, except when the wet bulb is frozen when it takes on the value 0.000720. It is important that accurate values of p are used otherwise relatively large errors may ensue. At relative humidities of 10%, for example, the assumption of p=1000 mb when the correct value is 950 mb gives rise to errors of about 20% in e, 2% in relative humidity and 2 °C in t_d.

The second variable to be calculated is t_d, which can be obtained from the calculated value of $e = e_s(t_d)$ by using the transposed Magnus equation for evaporation over water:

$$t_d = \left(\frac{239.0\,K}{17.38-K}\right)$$

where $K = \ln e - \ln 6.1070$.

The next step is the use of the Magnus formula for water to calculate the saturation vapour pressure at the dry-bulb temperature ($e_s(t)$). This enables the relative humidity to be calculated as $e/e_s(t)$.

These steps may be rationalized and summarized as follows:

(i) use the Magnus formula to calculate the saturation vapour pressure at t and t_w,

(ii) use the Regnault equation to calculate $e = e_s(t_d)$,

(iii) use the transposed Magnus equation to calculate t_d and

(iv) calculate the relative humidity as $e/e_s(t)$.

5. Estimation of the wet bulb from the dry bulb and dew-point

The estimation of t_w from t and t_d is a more complex procedure than obtaining t_d from t and t_w. This is essentially because the Regnault equation is expressed in terms of the depression of the wet bulb rather than the dew-point, and so can only be solved for the former by using an iterative procedure.

Less accurate estimates can, however, be obtained using an empirical approach which takes advantage of the fact that t_w lies between t and t_d. At low temperatures, when little water is available for evaporation, t_w is not much less than t, but at high temperatures, t_w is closer to t_d. In other words, the ratio of the wet-bulb depression to the dew-point depression always lies between zero and unity and increases with temperature. A convenient linear representation of this change is given by

$$\frac{t - t_w}{t - t_d} = 0.34 + 0.006\,(t + t_d).$$

For values of t from $-10\,°C$ to $+50\,°C$ and dew-point depressions up to $15\,°C$, errors in t_w obtained from this equation are always less than $0.3\,°C$. Improved accuracy could be obtained by introducing a non-linear dependence on $(t + t_d)$ to the right-hand side of the equation — the quoted relation produces a wet-bulb depression which exceeds the dew-point depression at temperatures above about 55 °C and is negative at temperatures below about $-25\,°C$. It may be more expedient, however, to solve the Regnault equation iteratively.

The Regnault equation may be rearranged as

$$e_s\,(t_w) + Apt_w = e_s\,(t_d) + Apt$$

Using the Magnus equation this becomes

$$6.107 \exp\left(\frac{at_w}{b + t_w}\right) + Apt_w = 6.107 \exp\left(\frac{at_d}{b + t_d}\right) + Apt.$$

All the terms involving t_w are now on the left-hand side. The right-hand side, involving terms in t and t_d, can be evaluated and set equal to a value C. We can then define:

$$F(t_w) = 6.107 \exp\left(\frac{at_w}{b + t_w}\right) + Apt_w - C \quad \ldots\ldots\ldots\ldots(6)$$

which takes on the value 0 for a correct solution of t_w. This can be obtained using the Newton–Raphson iterative approximation:

$${}_1t_w = {}_0t_w - F(t_w)/F'(t_w)$$

where ${}_0t_w$ and ${}_1t_w$ are the initial and improved estimates of t_w and $F'(t_w)$ is the first derivative of $F(t_w)$,

i.e. $$F'(t_w) = 6.107\,\frac{ab}{(b + t_w)^2}\exp\left(\frac{at_w}{b + t_w}\right) + Ap \quad \ldots\ldots\ldots(7)$$

The values of a, b, and A to be used in equations (6) and (7) are determined by the value of ${}_0t_w$.

The calculation of t_w from t and t_d may therefore be summarized as follows:

(i) make an initial estimate of ${}_0t_w = 0.5\,(t + t_d)$,

(ii) select the appropriate values of a, b, and A, i.e.

if ${}_0t_w \geqslant 0\,°C$	$A = 0.000799$	$a = 17.38$	$b = 239.0$,
if ${}_0t_w < 0\,°C$	$A = 0.000720$	$a = 22.44$	$b = 272.4$,

(iii) calculate $C = 6.107 \exp\left(\dfrac{17.38t_d}{239 + t_d}\right) + Apt,$

(iv) evaluate $F(t_w)$ and $F'(t_w)$ from equations (6) and (7),
(v) obtain an improved estimate ${}_1t_w = {}_0t_w - F(t_w)/F'(t_w)$,
(vi) test for convergence of the procedure, i.e. $({}_1t_w - {}_0t_w) < 0.005$.

If this condition is not met, set ${}_0t_w = {}_1t_w$ and repeat the procedure from step (ii). The criterion is usually satisfied in two or three cycles.

6. Meteorological Office procedures

The introduction of new WMO synoptic codes containing dry-bulb and dew-point temperatures to an accuracy of 0.1 °C created the possibility of increased automation by allowing the calculation of humidity parameters direct from the reported values of t and t_d. This opportunity has been taken by the Meteorological Office whose computer archives of observations from synoptic stations are now based on those contained in the coded messages instead of data keyed from manuscript tabulated returns. The wet-bulb temperatures so archived are no longer those measured, but are now obtained from an iterative solution of the Regnault equation. This peculiar circumstance could only be avoided by the introduction, as a national practice, of the reporting of t_w as well as t_d in the synoptic codes. The vapour pressure and relative humidity are then obtained from a simple application of the Magnus formula. For voluntary climatological stations, for which the wet-bulb temperatures are still received in manuscript form, humidity parameters are calculated using the procedures described in section 4.

The correct pressure to supply to the Regnault equation is that observed at the station, uncorrected for altitude. At voluntary climatological stations the pressure is seldom recorded and so a value of 1000 mb is used. At synoptic stations, users of humidity slide rules are instructed to assume a value of 1000 mb unless the pressure is less than 950 mb. Since the recovery of t_w from t_d is a reversal of a calculation made with a slide rule, the value of the pressure used in this procedure is also set to 1000 mb. For consistency, therefore, all algorithms used to calculate humidity parameters on pocket calculators should use a pressure of 1000 mb. If and when the preparation of synoptic messages is fully automated at all stations, then this could be replaced by the station level pressure.

7. Conclusions

The problems in calculating the various measures of humidity are caused by the difficulty in evaluating the saturation vapour pressure. This is related to departures from the ideal gas law and the variation of latent heat with temperature; these make the Clausius–Clapeyron equation difficult to integrate analytically. Semi-empirical equations due to Goff and Gratch (1945) are available for the accurate computation of saturation vapour pressure, but these are relatively costly to evaluate on a regular basis and for many purposes their accuracy is superfluous. Consequently many attempts have been made to devise simpler approximations providing only the accuracy required. In these attempts, simplicity has been interpreted as the speed and cost of evaluation on a large computer. In operational meteorology, however, there is the need to calculate dew-point from the wet-bulb temperature at synoptic observing stations, and this can be conveniently carried out on a programmable calculator. In this context algebraic simplicity is important and the brevity, reversibility and accuracy of the Magnus formula make it ideal for such a purpose. The Magnus formula is also perfectly suitable for implementation on a large computer and has been used for routine climatological calculations at the Meteorological Office for many years.

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Synoptic observations from Portland Bill

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Introduction

Synoptic observations from the keepers of Portland Bill lighthouse recommenced on 18 July 1984 after a break of 15 years, during which period observations were supplied by the Coastguard Service. This article reviews the history of the observing stations at Portland Bill since 1899, and describes some of the problems encountered. Many scores of thousands of useful observations have been made with both forecaster and mariner benefiting from the regular reports, particularly those of fog and wind conditions over the English Channel.

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Station history

Synoptic weather reports made for the Meteorological Office by the lighthouse keepers commenced at Portland Bill lighthouse from an enclosure 54 metres (m) above mean sea level (amsl) on 3 June 1899. The establishment of the weather station was an economy measure which allowed the closure of the stations at Prawle Point and Hurst Castle (Fig. 1).

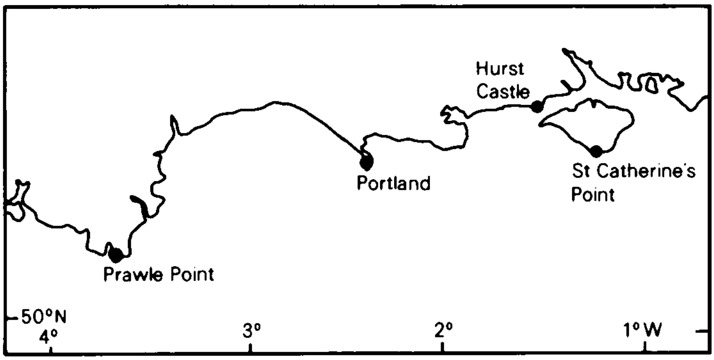

Figure 1. Places mentioned in the text.

In 1906 the meteorological instruments were transferred to the new lighthouse near the Bill (enclosure 6 m amsl — see Figs 2 and 5). No record is available of anything eventful in the very early years, apart from the breakage of both barometers on 1 June 1909 when the wall fitting collapsed.

The thermometer screen was re-sited near the base of the lighthouse in 1913 (Fig. 3). Guy wires were used to brace the screen against strong winds.

The early 1920s saw an increasing number of day trippers, and in the summer of 1923 a motor car ran into the 8-inch rain-gauge. This was no mean feat considering the size of the rain-gauge and the paucity of cars in those days, although the incident was repeated several times the following year. The 1924 inspector reported that holiday makers used the rain-gauge as a foot scraper. Cars and holiday makers were subsequently kept at bay by a 3 ft 6 in paling fence, 15 ft radius from the rain-gauge. However, objections by the Court Leet of the Manor of Portland led to the removal of the fencing in April 1926, followed by car damage to the rain-gauge within three months. A more secure site for the rain-gauge was found in October 1926 on land owned jointly by the Crown and commoners of Portland, for which the Air Ministry paid a yearly rental of 2s.6d. (12½p) to the Crown and 7s.6d. (37½p) to the commoners. The fence was re-erected. The gauge gave low values compared to readings at Portland Royal Naval Air Station (RNAS) and a turf wall was constructed inside the paling fence. This made no noticeable difference and the rain-gauge problems continued throughout the life of the station.

In October 1945 the lighthouse was opened to the public. A car park was provided by the council, and six cafés catered for the visitors. The following summer a harassed inspector, Mr (now Professor) Hubert H. Lamb, found much difficulty in gaining the keepers' attention because of the number of visitors — 1600 to 1800 a day. The magnifying glass provided to assist in the reading of the mercurial barometers soon became a memento for one of them.

In 1947 a shortage of regular light-keepers caused difficulties at the station. Naval ratings were brought in to act as relief light-keepers and, to their astonishment, as auxiliary meteorological observers.

Figure 2. Portland Bill new lighthouse looking east, from an old photograph.

Figure 3. 22 May 1913. Light-keeper Withers (left) and Assistant Light-keeper Ball (right), the observers, by the thermometer screen. Note single-louvered door and mounted thermometers.

The strategic position of the station was given further recognition from 22 April 1956, when Portland Bill's observations were first included in the BBC Shipping Bulletins.

An electrical anemograph system was installed on the lantern roof in March 1963 (Fig. 4). Before this, wind directions had been obtained from the Trinity House vane on top of the lighthouse, and wind speeds estimated.

In 1966 Meteorological Office requirements for hourly full synoptic reports could not be met by the keepers, so the assistance of the nearby coastguards as auxiliary observers was sought. Hourly full synoptic observations by HM Coastguards commenced from an enclosure (53 m amsl) at their look-out (Fig. 5) on 1 January 1967. The thermometer screen and instrumentation were supplied by the Meteorological Office. An anemometer system was erected on a 40 ft lattice tower 150 yards to the east of the meteorological enclosure with an anemograph recorder at the look-out. Overlap observations restricted to 09 and 21 GMT continued at the lighthouse until 31 March 1968, and anemograph records until 31 March 1969 after which the recorder was withdrawn although the rest of the anemometer system remained.

Figure 4. Electrical anemometer head and vane on top of the lighthouse. The exterior ladder allows access for servicing.

In 1979 it was noticed that the number of breakages of maximum and minimum thermometers was above average. The breakages were attributed to buffeting in strong winds which caused the instruments to move horizontally along their rests until they slipped from the end of the supports and fell through the slatted floor of the thermometer screen to the ground below. Rubber grommets were supplied to slip on to the ends of the thermometers to prevent them from moving along their rests, and these proved successful.

Reorganization of the Coastguard Service meant that from 1 January 1984 the look-out was not manned regularly at night and observations were lost. An appeal for the light-keepers' assistance was readily answered by three of the keepers who agreed to provide three-hourly full synoptic reports when on duty. A thermometer screen has been provided in the lighthouse compound, but there remains no satisfactory exposure for a rain-gauge.

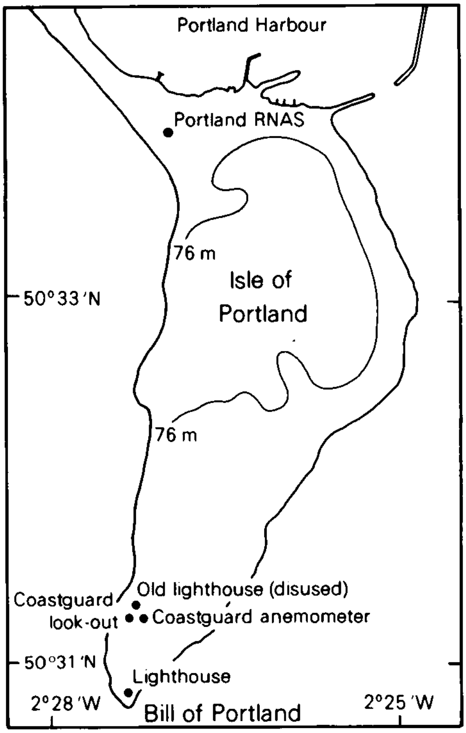

Figure 5. Reporting stations on Portland.

The wheel has thus turned full circle with an Auxiliary Reporting Station re-established at the lighthouse. The original anemometer on the lighthouse lantern is once more used for meteorological observations, although the coastguards still maintain the anemograph record from the anemometer provided for their reports. However, bulletins broadcast for mariners by the BBC now include reports from St Catherine's Point, rather than Portland, in view of their hourly frequency throughout the 24 hours. The Portland reporting program is now:

03855	Portland Bill/Coastguard 53 m amsl	05 (summer), 06 to 17 hourly, 18 (winter) GMT plus other hours when the look-out is manned for coastguard casualty working or bad weather watch. Climatological and anemograph records.
03856	Portland Bill (lighthouse site) 11 m amsl	00, 03, 06, 09, 12, 15, 18, 21 GMT when a co-operating keeper is on duty.

Observations are also made from Portland RNAS (03858) 4 m amsl, when flying is taking place, although the site at this station has a mainly northerly aspect.

Letter to the Editor

Comments on 'Large hail over north-west England, 7 July 1983' (by L. Dent and G. Monk, *Meteorol Mag*, **113**, 1984, 249–263).

(i) You may be interested to know that on 7 June 1983 at 1420 GMT an exceptional thunderstorm occurred here at Chivenor which produced hailstones measuring ½ inch in diameter. The visibility was reduced to 90 metres and a gust of 47 knots was recorded with the temperature falling to 6 °C.

(ii) This storm occurred some 2½ hours earlier than those noted in the above article.

(iii) Cumulonimbus tops were reported to over 40 000 feet.

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Notes and news

75 years ago

The following extract is taken from *Symons's Meteorological Magazine*, February 1910, **45**, 6–7.

THE GREAT PARIS FLOOD OF 1910

It has been found impossible to obtain data as to the meteorological conditions which led to the widespread floods in France during the last ten days of January in time for a satisfactory account of the disaster in the present issue. We hope to deal with the subject at a later date. It is curious that in the many columns devoted daily by the London press to the spread of the floods, the amount of the rainfall has not been definitely stated, and it remains for the present a matter of speculation. Heavy and continuous rains had fallen for some days before January 21st, and rivers were in flood all over France, but public attention was naturally concentrated on the state of matters in Paris, where damage was done on a scale unequalled in human memory. Indeed, the flood may prove to have been the highest on record. At the Pont Royal the height of floods in the Seine has been recorded for many years, and we quote, on the authority of the Paris correspondent of *The Times*, the following extreme measurements of the height of the river at this point—in 1615 over 32 feet, in 1802 nearly 30 feet, in 1876 about 26 feet, and on January 29th, 1910, over 31 feet. The height of the great flood of 1658 at the Pont de la Tournelle was 29 feet, and that of 1876 was about 21 feet. If we may take 5 feet as the difference in level between the two datum marks it would appear that the present flood fell slightly short of that of 1658.

If data of scientific value have been scanty in the English press, descriptions of the extent of the inundations on both sides of the Seine and of the damage done have been bewilderingly abundant. While the waters were rising traffic was stopped over many of the bridges; but although only a few inches of the central arch of the Pont d'Alma remained above water, none of the bridges collapsed, and traffic was resumed before the flood abated. It is said that 200,000 people were seriously affected by the flood, being either driven from their homes or thrown out of work in consequence of the stoppage of factories or the closing of shops. The low-lying parts of the underground railway system were completely filled with water, and several of the great railway termini were cut off and closed for traffic. This was the case with the Gare de Lyon and the Gare d'Orleans; the traffic to the Mediterranean coast through Paris was suspended, and Mr. Asquith made his way to Nice after the General Election by a roundabout route through Bâle and Genoa. More than 15,000 telephone subscribers in Paris were cut off, and most of the telegraph wires out of Paris also failed for a time. Water invaded many public buildings, and divers were employed to rescue the archives of the Palais de Justice from under 8 feet of water. The Louvre was seriously threatened, and the spread of the flood was only controlled by the exertions of troops with bags of cement, sand and stones torn from the pavements, which were used to form barricades against the water. Sailors and boats from the Navy were hurried to Paris to save the inhabitants of the inundated streets, and there was practically no loss of life, thanks to the admirable organization shown in dealing with the situation. The extent of the damage is unknown at the time of writing, and must amount to many million pounds. The chief danger is to the stability of buildings by subsidences in the streets as the water goes down, and to health by the accumulation of sewage from the burst sewers. It is probably not too much to say that the flood has been more widespread and productive of distress than any previously known in Europe, though, considering its extent, it has been singularly free from loss of life.

The complete disorganization of the life of a great city in consequence of some unfortunate combination of meteorological conditions is very rarely seen, and it has been pointed out in Paris that the area inundated would have been far less and the resulting suffering not nearly so great if the quay-walls along the river had not been so frequently cut in recent years for facilitating traffic, and if the system of underground railways had not been so greatly developed.

Reviews

Plants and microclimate. A quantitative approach to environmental plant physiology, by Hamlyn G. Jones. 160 mm × 235 mm, pp. xviii + 323, *illus.* Cambridge University Press, Cambridge, London, New York, New Rochelle, Melbourne, Sydney, 1983. Price £27.50, £12.50 (paperback).

The investigation of the relations between plant growth and environment is a topic for both meteorologists and plant physiologists. Most meteorologists who work in this field will probably have made their way by means of books and papers written by physical scientists (e.g. *Principles of environmental physics* by J. Monteith). Plant physiologists, however, working from 'their side' have also made significant progress and here is a book by a plant physiologist which, while firmly rooted in plant physiology, makes a useful contribution to the application of physical principles to the solution of plant growth and ecological problems. The author compiled the book from courses in environmental physiology given to undergraduates, but the treatment is not superficial, with some good discussion of advanced topics.

After an introductory chapter on modelling and experimentation the book is divided into 11 other chapters. Three of these cover the physics of radiation, heat, mass and momentum transfer and also energy balance and evaporation. Much of this material will be familiar to meteorologists, but there is a strong emphasis on molecular transport processes since this is the main agent of diffusion within plants. The chapter on radiation emphasizes the importance of the number of quanta of photosynthetically active radiation rather than total energy in the study of radiation within plant canopies. On page 19 it is unfortunate that the regression coefficients in the relation between global short-wave irradiance and sunshine hours use work done by L. P. Smith nearly 20 years ago and not the more recent appraisal by J. P. Cowley (1978) published in the *Meteorological Magazine* (The distribution over Great Britain of global solar irradiation on a horizontal surface, *Meteorol Mag,* **107**, 357–373).

Similarly, on page 96, L. P. Smith's grassland transpiration model is quoted as still being 'widely used in the UK'. This is not true, and even in 1982 when the book was produced the MORECS system had been in operational use for several years and written up as *Hydrological Memorandum No. 45* in 1981 (The

Meteorological Office rainfall and evaporation calculation system: MORECS (July 1981), by N. Thompson, I. A. Barrie and M. Ayles).

The remaining eight chapters form the bulk of the book and this is where the author takes us right inside plants to explain their growth reactions. There are chapters on the following topics: plant– water relations, stomata, photosynthesis and respiration, plant morphogenesis, temperature, drought, wind altitude and CO_2, and finally a chapter on yield improvement. The water relations of plants are described using water potential, a concept unfamiliar to meteorologists, but here we have a clear account to enable us to read plant physiological literature which uses these ideas. All the factors which could possibly affect stomatal resistance are set out and there is a critical appraisal of the methods currently in use to measure it. The photosynthesis chapter is clearly set out though some of the chemistry is beyond this reviewer. The C_3, C_4, and CAM Photosynthetic pathways are clearly distinguished and there is a detailed discussion of resistances to CO_2 diffusion. Some parts of this chapter would be hard going for most meteorologists. The chapter on plant morphogenesis would also have material unfamiliar to meteorologists, but here the mysteries of phytochrome, photoperiodism and nastic movements are explained. The chapters on temperature and drought discuss how plants react to these environmental factors in terms of photosynthesis, injury and hardening. There is a strong ecological thread to these chapters which tries to explain some of the observed facts of the distribution and natural selection of plants for and within specific environments. There is an interesting section on water use efficiency (i.e. the ratio of net assimilation to water loss) which leads to a section on the modelling of optimum stomatal behaviour. The chapter on wind, altitude and CO_2 is rather more sketchy than the others, but still gives much useful information. A treatment of lodging is given which fails to mention that the main effect of rain on cereals is to increase the weight of the ear and so increase the turning moment about the stem base when the wind blows. The last chapter attempts to conclude what the ideal crop plant ought to be like using some of the ideas from the rest of the book. There are ten appendices giving derivations of some of the formulae used in the text, and then 20 pages of references many of which are from the 1970s and 80s. The index is comprehensive.

For a book of this size and content it is remarkably free from errors. I could find only two: Figure 7.18 where one symbol in the key does not match the diagram and the other in Figure 12.5 where wheat yields of up to 70 t/ha are plotted.

The level of knowledge expected of the reader is variable. Some of the chemistry required has already been mentioned and while torque, for example, is explained on page 242, the Gibb's free energy is mentioned on page 64 without more ado. However, these are minor complaints about an excellent book which all agricultural meteorologists would benefit from using.

M. N. Hough

Milankovitch and climate, edited by A. Berger, J. Imbrie, J. Hays, G. Kukla and B. Saltzman. 2 Vols 155 × 235 mm, pp. Part I xxxiv + 510, Part II ix + 510, *illus.* D. Reidel Publishing Company, Dordrecht, Boston, Lancaster, 1984. Price Dfl 310, US $117.00.

This pair of substantial volumes records most of the papers presented at an international symposium held at the Lamont-Doherty Geological Observatory, Palisades, New York, in December 1982. The aim of the meeting was to consider and refine the theory of Milankovitch (1941) that the Pleistocene ice ages were triggered by variations of the Earth's orbit around the sun.

A systematic account is given of all aspects of current research into the mechanisms underlying the inception, maintenance and termination of the Pleistocene ice ages. Subjects discussed include methods of computing more accurately the orbital variations themselves; geological evidence for Pleistocene and earlier long-term climatic variations with the periodicities predicted by Milankovitch (approximately 19 000, 23 000, 41 000 and 100 000 years); and modelling long-term climatic variation which may occur in response to orbitally-induced forcing.

All sections of the work convey an atmosphere of new discovery and insight, combined with the realization that many improvements and further discoveries remain to be made. For this reason the papers will rapidly become dated, though they will retain historical interest as examples of scientific methods, co-operation, and progress — V. Milankovitch's memoir of his father is already interesting material for the historian.

A key paper is that of Imbrie *et al.* who use records of the oxygen isotope O^{18} from marine cores to leave little room for doubt that variations in the geometry of the Earth's orbit are the main cause of the succession of late Pleistocene ice ages. The oceans are richer in O^{18} when global ice volume is large, because of preferential evaporation and condensation. Given radiometric and magnetic dating, the cores are found to be consistent with predictions using the 19 000- and 23 000-year precessional cycles and the 41 000-year obliquity cycle, assuming a 17 000-year time-constant for the Earth's response, and allowing phase-tuning to refine the dating. The criticism of possible over-tuning or over-fitting is largely answered by the good fit to the 100 000-year eccentricity cycle which was not included in the phase-tuning of the cores.

By contrast, the uncertainty of many deductions about Pleistocene climate is highlighted by the paper of Janecek and Rea who deduce that glacial periods were moist in middle and low latitudes, in contrast to the aridity claimed by other authors. The analysis of Pacific aeolian sedimentation rates and grain sizes in their paper clearly shows the main Milankovitch cycles, but although the obliquity and precessional cycles in grain size indicate stronger winds during glacials, the eccentricity cycle in grain size indicates the reverse. To explain this the authors had to invoke a change of latitude of the mid-latitude westerlies in association with the eccentricity cycle in glaciation, without invoking a similar change of latitude in response to the obliquity and precessional modulations of glaciation. This is very reasonable, in view of the lesser size of these modulations compared with the apparent imprint of eccentricity on glaciation, but is nevertheless still speculation.

Pestiaux and Berger present a particularly illuminating and useful review of the variety of spectral analysis methods now available. An appreciation of, for example, the somewhat different results which different spectral analysis methods give for the same data will enable the reader to assess more critically the results presented in this (and many other) fields. This paper may not date as rapidly as some of the others.

The paper of Ruddiman and McIntyre is recommended as a thought-provoking review article which considers some possible feedbacks within the atmosphere–ocean–cryosphere system in the context of the precessional and obliquity signals during the Pleistocene. Broecker's invocation of atmospheric carbon dioxide changes as a trigger for rapid deglaciations contrasts markedly with Pollard's ice-calving mechanism and shows that the field in this area remains wide open for debate.

A few of the papers betray hasty proof-reading, and there are some printer's errors. Pages 661 and 666 are interchanged, and the printing reproduction fails to do justice to the diagram on page 66. The authors of the Meteorological Office's contribution on page 721 are omitted from the authors' index. That particular paper is a good example of one which is already out of date (see *Nature* 23 August 1984).

At the end of Volume II there is a summary chapter drawn up at a small subsequent workshop. Although useful it makes less inspiring reading than many of the papers. The recommendations for future work demonstrate how much remains to be done. Simulations of 100 000 years with a realistic fully coupled general circulation model of the atmosphere–ocean–cryosphere system are still a somewhat distant target!

D. E. Parker

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