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Retirement of Mr D.E. Miller

Mr Derek Miller, Deputy Director (Communications and Computing), took retirement from the Meteorological Office at the early age of fifty years on 31 July 1987 after a career of twenty-nine years in the Office, most of which were spent in the area of high atmospheric research and in the development and use of rockets and satellite instrumentation to make measurements of the atmosphere.

Mr Miller was educated at Magdalen College School, Brackley (Northants) and at Wadham College Oxford, where he graduated with a first class Honours (BA) in Natural Sciences and won the Scott Prize for Physics in 1958. He joined the Office directly from Oxford, and took the usual Scientific Officer course at Stanmore where he was immediately recognized as one of those rare people who combine a first-class academic ability with a very practical approach to the problems of applied meteorology. Thus, after a short spell of forecasting experience at Wyton and London (Heathrow) Airport in 1959, he was posted into research with the High Atmosphere Research Unit which, at that time, was located at Kew Observatory under Dr K.H. Stewart.

Thus began a direct relationship with high-altitude and satellite meteorology which lasted up until 1985. At Kew he became deeply involved in the design and development of the instrumentation required to make observations of stratospheric ozone from Skylark rockets and from the Anglo-American satellite Ariel 2.

In October 1962 he was promoted to Senior Scientific Officer and continued to work in the same unit, which was now part of the Satellite Meteorology Branch (Met O 19), under Dr Stewart and Dr R. Frith. As the UK representative at the National Aeronautics and Space Administration in Washington, during tests of the equipment for the Ariel 2 satellite, his ability and personality earned great respect in the USA.

By 1965 the team had moved to Bracknell, and the Ariel 2 satellite ozone experiment involved him in programming with the KDF 9 computer, and a joint project between the Office and Dr J.T. Houghton's group at Oxford University on the measurement of molecular oxygen airglow in the infra-red was started.

Mr Miller was promoted to Principal Scientific Officer in 1967, staying in Met O 19, and that year was notable for his paper in the *Proceedings of the Royal Society* on measurements of stratospheric attenuation in the near ultraviolet. His work then expanded to cover the preliminary design for the instruments to measure molecular oxygen and ozone on the Nimbus F spacecraft, and further rocket work in Europe with the European Space Research Organization (ESRO), and in Canada and the USA.

From 1967 to 1976 he continued to be deeply involved in the development of new methods of observation, sensors and instruments on meteorological satellites and rockets. Papers were produced on the results of Skua rocket flights, and on an analysis of the correlation between air density and magnetic disturbances deduced from the change in the Ariel 2 satellite spin rate. He became a member (later the Chairman) of the ESRO Working Group on the Ground Facilities required for the new Meteosat Project, Chairman on the Working Group for the Meteorological Information Extraction Centre (MIEC) Software Development, and also was closely involved in the scientific development and contract monitoring of the Stratospheric Temperature Sounder Unit for the TIROS-N satellite. He became well known in European circles and was much sought after for his knowledge and experience.

In 1976 Derek Miller was promoted to Senior Principal Scientific Officer and became the Assistant Director of the Satellite Meteorology Branch. In this post he was able to use his administrative abilities and powers of scientific leadership. Much of his time was also devoted to national and international committee work and to consultancies with the European Space Agency (ESA) concerning the geostationary satellite Meteosat. He was a member of the Science Research Council Solar System Working Group and continued as Chairman of the *ad hoc* Group on Satellite Data Reception for Meteosat/ESA.

In 1982 he was again promoted, to Deputy Chief Scientific Officer (now Grade 5) as Deputy Director of Physical Research. In this post, looking after the administration and management of a number of branches ranging from Boundary Layer Research to the Meteorological Office Radar Research Laboratory at Malvern, his abilities made him a tower of strength, and he continued his interests in space meteorology as Chairman of an inter-departmental team charged with defining a costed national remote-sensing programme in support of the Earth Resources Satellite, ERS-1, and through working on the Natural Environment Research Council, the Science and Engineering Research Council and Royal Society committees on meteorological research in the universities and in industry.

In 1985 Mr Miller moved to the new area of Deputy Director (Communications and Computing) where he was able to exercise his administrative and organizational talents in a new field. He continued to set himself the highest standards, and his insight and judgement have been very evident, in particular in a recent study on the future of the Man–machine Interface in Forecasting. He has worked particularly hard on the new plans for the replacement of the Cyber 205 supercomputer, the Weather Information

System and the new satellite processing system AUTOSAT-2. All of these are going well despite the fact that he found the frustrations and delays inherent in the projects difficult to balance with his self-set standards of excellence.

Mr Miller is a man of energy and drive, and I cannot imagine that he will be idle for very long in his retired state. I shall miss his sound judgement and attention to detail in matters both large and small, and not least for his wise advice as Chairman of the Accommodation Committee over the past two years. I am sure that all his friends and colleagues will join me in wishing Derek a long and happy retirement, and I have no doubt we shall be meeting him again in matters of meteorological significance in the future.

D.N. Axford

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An introduction to the parametrization of land-surface processes Part II. Soil heat conduction and surface hydrology*

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Summary

This paper concludes an introduction to the subgrid-scale land-surface processes which, it is generally acknowledged, need to be included by parametrization in three-dimensional numerical models for studying climate and climate change, and for numerical weather prediction. The discussion is restricted in the main to the relatively simple case of non-vegetated land surfaces.

Part I (Carson 1987) described the general boundary conditions for momentum transfer and the balance equations for energy and mass (moisture) transfer at a bare-soil surface. The surface radiative properties and fluxes, and the physical character and the parametrization of the surface turbulent exchanges, were considered.

Part II concentrates mainly on soil heat conduction and the land-surface temperature, and surface hydrology and the soil water budget. It concludes with a brief discussion of some of the particular problems associated with snow-covered, non-vegetated, land surfaces.

1. Introduction

In Part I of this introduction to land-surface processes (Carson 1987) the balance equations for energy and mass (moisture) transfer at a bare-soil surface were described. Also the surface radiative properties and fluxes were considered along with the physical character and parametrization of the surface turbulent exchanges. Here emphasis will be placed on two topics:

(a) soil heat conduction and the prediction of the land-surface temperature using the surface energy balance (section 2);

(b) surface hydrology and the prediction of the soil moisture content using the moisture balance equation (section 3).

The way in which snow-covered surfaces are taken into account will also be considered (section 4).

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There is a need to study heat conduction in the soil to provide a sound physical basis for evaluating the surface temperature. In the same way it is necessary to understand the dynamics which govern the movement of water in the soil so that changes in soil moisture content can be modelled. Although the concepts of surface temperature and soil moisture content are discussed separately, it is worth emphasizing the strong interactive coupling between the thermal and hydrological properties and processes in the soil. Not only does the evaporation rate appear explicitly in both the balance equations for energy and mass but also most of the other surface fluxes (including the momentum flux) depend, to varying degrees, on the surface temperature and soil moisture content. Indeed, in a model with both interactive surface hydrology and interactive land-surface temperature, the value of the soil moisture content has an important bearing on the surface temperature and vice versa.

2. Soil heat conduction and the land-surface temperature

The energy flux balance at a bare-soil surface may be expressed as

$$G_0 = R_N - H - Q \qquad \dots \dots \dots \dots \dots \dots \quad (1)$$

where R_N is the net radiative flux, H is the turbulent sensible-heat flux, $Q = L_e E$ is the latent-heat flux due to surface evaporation (E is the surface evaporation rate and L_e is the latent heat of evaporation) and G_0 is the flux of heat into the soil.

If the aim was solely to evaluate G_0, then use of the surface energy balance as given by equation (1) would be a legitimate method of obtaining such an estimate. Indeed, in principle, the energy balance method can be invoked to estimate any one of the terms in equation (1) if all the other terms are known by some other means.

The more direct, microphysical approach to understanding the soil heat flux term, G_0, is through the study of heat transfer in the soil itself, a process which is predominantly that of heat conduction. In general, G_0 will depend in a complicated way on the soil's thermal properties which in turn depend on, for example, the type of surface, the type of soil and whether it is wet, dry, frozen or snow-covered, and whether it is bare soil or vegetation. In simple, general terms a thin surface layer of the soil stores heat during the day (strictly, from equation (1), when $R_N > H + Q$, i.e. G_0 is positive) and acts as a source of heat energy to the surface at night (strictly, when $R_N - H - Q < 0$, i.e. G_0 is negative). On longer, seasonal and annual time-scales deeper soil layers act as a reservoir of heat which may be replenished during warm seasons and depleted during cold seasons.

Good estimates of the detailed behaviour of G_0 throughout the day and throughout the year are now recognized as important for inclusion in numerical weather prediction models (NWPMs), which attempt to forecast the characteristic diurnal cycle of land-surface temperatures, and also in climate models which need to simulate realistically and interactively the heat-storage properties of the soil over periods ranging from less than a day to at least several years.

Implicit in a knowledge of heat transfer through the soil is a knowledge of the soil temperature profile with depth. In particular, the land-surface temperature, T_0, features in each of the terms in equation (1) and it is now the common practice in atmospheric general circulation models (AGCMs) and NWPMs to invoke the surface energy balance as a diagnostic relation or prognostic equation for evaluating T_0. The variety of techniques commonly used in such models for representing G_0 in the surface energy balance has already been reviewed fairly comprehensively by, for example, Bhumralkar (1975), Deardorff (1978) and Carson (1982, 1986). In order to illustrate the relationships between soil heat flux, soil temperature profile and the thermal properties of the soil, I shall restrict my discussion to those methods which rely on a knowledge of heat conduction in the soil and invoke either simple, one-dimensional, analytical models or attempt to model explicitly the soil heat transfer in a multi-layer soil model.

2.1 *Heat transfer in a semi-infinite homogeneous soil*

Most parametrizations of G_0 are now based on considerations of heat conduction and conservation in the soil. The problem is usually simplified by assuming a semi-infinite, spatially homogeneous soil layer with no horizontal heat transfer and no melting or freezing within it. This restricted and idealized one-dimensional problem is governed by the following two equations:

(a) *The soil heat conservation equation*

$$\frac{\partial T_g}{\partial t} = -\frac{1}{C}\,\frac{\partial G}{\partial z_g} \qquad \ldots \ldots \ldots \ldots \ldots \quad (2)$$

where T_g is the soil (ground) temperature, G is the soil heat flux, C is the volumetric heat capacity of the soil (units J m^{-3} K^{-1}), $z_g = -z$ is the vertical coordinate in the soil layer and t is time.

(b) *The flux–gradient relation for heat conduction*

$$G = -\lambda\,\frac{\partial T_g}{\partial z_g} \qquad \ldots \ldots \ldots \ldots \ldots \quad (3)$$

where λ is the thermal conductivity of the soil (units W m^{-1} K^{-1}).

Substitution of equation (3) into equation (2), with the assumption of homogeneity, yields the one-dimensional equation for the conduction of heat in the soil

$$\frac{\partial T_g}{\partial t} = \kappa\,\frac{\partial^2 T_g}{\partial z_g^2} \qquad \ldots \ldots \ldots \ldots \ldots \quad (4)$$

where κ is the thermal diffusivity of the soil (units m^2 s^{-1}). The thermal diffusivity is related to the other soil parameters by $\kappa = \lambda/C = \lambda/\rho_g c_g$ where ρ_g is the uniform soil density and c_g is the specific heat capacity (units J kg^{-1} K^{-1}).

The definitions and characteristics of the soil thermal properties C, λ, κ and c_g can be found in standard textbooks such as Geiger (1966), Sellers (1965), Oke (1978) and Rosenberg *et al.* (1983). The values in Table I come from Oke (1978) and illustrate the typical magnitudes of these terms for a few simple soil types (and for snow); an indication is also given of the sensitivity of the parameters to how wet or dry the soil is.

It is standard practice to use equations (1), (2) and (3) to produce a prognostic equation for T_0 (usually assumed equivalent to the soil surface temperature T_{g0}). The simplest approaches of this kind introduce the concept of an effective depth of soil, D, and an effective surface thermal capacity $C_{\mathrm{eff}} = CD = \rho_g c_g D$ defined such that

$$G_0 = C_{\mathrm{eff}}\,\frac{\partial T_{g0}}{\partial t} = CD\,\frac{\partial T_0}{\partial t}\,. \qquad \ldots \ldots \ldots \ldots \ldots \quad (5)$$

Note that expressing G_0 in this way does not imply a priori that T_0 has been replaced by and equated to the mean temperature throughout the shallow surface soil layer of depth D, with neglect of conduction of heat to or from lower soil layers. Rather it is an attempt, if D can be determined sensibly, to parametrize the changing heat content of the whole soil layer by using the analogue of a single shallow layer of known thermal capacity, C_{eff}, which is fully insulated at its lower boundary and whose uniform temperature, T_0, increases or decreases uniformly in response to G_0.

Substituting equation (5) into equation (1) and remembering the assumption that $T_0 = T_{g0}$ gives

$$C_{\mathrm{eff}}\,\frac{\partial T_0}{\partial t} = R_N - H - Q \qquad \ldots \ldots \ldots \ldots \ldots \quad (6)$$

Table I. *Thermal properties of natural materials (from Oke 1978). For explanation of symbols see text.*

Material	Remarks	ρ_g kg m^{-3} × 10^3	c_g J kg^{-1} K^{-1} × 10^3	C J m^{-3} K^{-1} × 10^6	λ W m^{-1} K^{-1}	κ m^2 s^{-1} × 10^{-6}
Sandy soil (40% pore space)	Dry	1.60	0.80	1.28	0.30	0.24
	Saturated	2.00	1.48	2.96	2.20	0.74
Clay soil (40% pore space)	Dry	1.60	0.89	1.42	0.25	0.18
	Saturated	2.00	1.55	3.10	1.58	0.51
Peat soil (80% pore space)	Dry	0.30	1.92	0.58	0.06	0.10
	Saturated	1.10	3.65	4.02	0.50	0.12
Snow	Fresh	0.10	2.09	0.21	0.08	0.10
	Old	0.48	2.09	0.84	0.42	0.40

which can be solved for T_0 provided the radiative and turbulent fluxes at the surface are known along with C_{eff}. Many AGCMs and NWPMs contain rather arbitrary and empirical selections of C_{eff} (see, for example, Carson 1982).

The effective depth can be defined more formally by considering the soil heat conservation equation. If $G \rightarrow 0$ as $z_g \rightarrow \infty$ then equation (2) can be integrated to give

$$G_0 = C\int_0^\infty \frac{\partial T_g}{\partial t}\, dz_g . \qquad \dots\ \dots\ \dots\ \dots\ \dots\ \dots\ (7)$$

Substituting equation (7) into equation (5) allows D to be defined in a strictly mathematical sense as

$$D = \left(\int_0^\infty \frac{\partial T_g}{\partial t}\, dz_g\right) \Big/ \frac{\partial T_0}{\partial t} . \qquad \dots\ \dots\ \dots\ \dots\ \dots\ (8)$$

2.2 *An analytical approach*

The problem of determining D can be overcome by appealing to the theory of heat transfer in a semi-infinite homogeneous medium when the surface is heated in a simple periodic manner (as discussed, for example, by Sellers 1965).

Assume that the surface temperature is given by

$$T_0 = T_g(0, t) = \hat{T}_g + a_0 \sin \omega t \qquad \dots\ \dots\ \dots\ \dots\ (9)$$

where ω is the angular frequency of oscillation, $\hat{T}_g$ is the mean soil temperature (over the period $P = 2\pi/\omega$) which is assumed to be the same at all depths, and a_0 is the amplitude of the surface wave. It can then be shown that the solution of equation (4) is

$$T_g(z_g, t) = \hat{T}_g + a_0 \exp(-z_g/\delta)\sin(\omega t - z_g/\delta) \qquad \dots\ \dots\ \dots\ (10)$$

where $\delta = (\kappa P/\pi)^{½}$ is the e-folding depth of the temperature wave of period P, i.e. it is the depth where the amplitude of the oscillation is reduced to $1/e \approx 0.37$ times its surface value.

The effective depth, D, corresponding to this soil temperature profile can be found by using equations (9) and (10) in equation (8) to give

$$D = \frac{\delta}{\sqrt{2}} \frac{\sin(\omega t + \pi/4)}{\cos \omega t} . \qquad \ldots \ldots \ldots \ldots \ldots (11)$$

Clearly D is not only a function of the thermal diffusivity of the soil and the period of the temperature forcing at the surface but also varies with time. Substituting for D from equation (11) into equation (5) gives a prognostic equation for T_0 which may be written as

$$\frac{\partial T_0}{\partial t} = \frac{2G_0}{C\delta} - \frac{2\pi}{P}(T_0 - \hat{T}_g) . \qquad \ldots \ldots \ldots \ldots \ldots (12)$$

This expression was first proposed, by a different argument, by Bhumralkar (1975) and has come to be referred to as the 'force-restore method', a term introduced by Deardorff (1978).

The period of the diurnal temperature oscillation is normally used in equation (12) as that appropriate for determining the thermal capacity of the effective surface layer. It is also necessary to specify $\hat{T}_g$. For short periods of a few days $\hat{T}_g$ may be fixed or diagnosed, but for longer periods of integration (e.g. in climate modelling) it needs to be determined prognostically. Deardorff (1978) has suggested a second prognostic equation for $\hat{T}_g$ analogous to equation (12) but with the appropriate effective depth determined by the e-folding depth of the annual temperature wave. However, extensions of this analytically based method (e.g. Deardorff 1978 and Carson 1982) soon become analogous to the more elaborate parametrization schemes which explicitly model the temperature profile through several soil layers.

2.3 *Multi-layer soil models*

The somewhat idealized analytical assumptions underlying the force-restore method and other simpler parametrization schemes can be avoided in principle by explicit modelling of the soil temperature profile and soil heat conduction with a multi-layer soil model of specified depth and with appropriate vertical resolution and boundary conditions. One approach would be to invoke equation (3) to evaluate G_0 explicitly from the modelled soil temperature profile such that

$$G_0 = \left(-\lambda \frac{\partial T_g}{\partial z_g} \right)_{z_g = 0} .$$

With this representaton of G_0, equation (1) could be solved diagnostically for T_0.

An alternative multi-layer approach is represented schematically in Fig. 1 for a three-layer soil temperature model. T_0 is represented by the mean temperature of a very thin surface layer of depth δ_0. Therefore, using simple finite differences, equation (2) becomes

$$\frac{\partial T_0}{\partial t} = \frac{G_0 - G_1}{C\delta_0} . \qquad \ldots \ldots \ldots \ldots \ldots \ldots (13)$$

Here G_0 is the net imbalance of the radiative and turbulent fluxes at the surface as given by equation (1) and G_1 is the soil heat flux into the next layer down, which is determined from the flux–gradient relationship by writing equation (3) in finite difference form

$$G_1 = \frac{\lambda(T_0 - T_{g1})}{½(\delta_0 + \delta_1)} . \qquad \ldots \ldots \ldots \ldots \ldots \ldots (14)$$

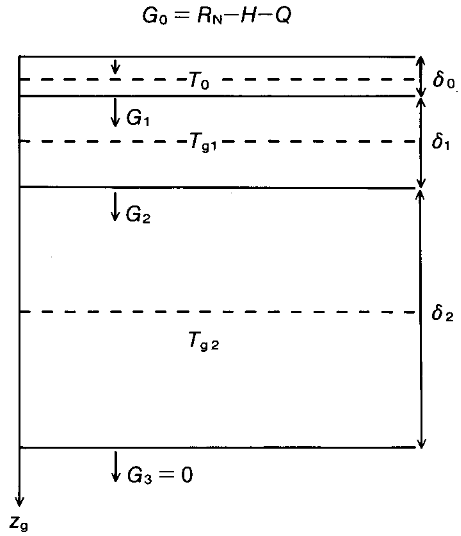

Figure 1. Schematic representation of a three-layer soil temperature finite difference model. For explanation of symbols see text.

Therefore equations (13) and (14) give

$$\frac{\partial T_0}{\partial t} = \frac{G_0}{C\delta_0} + \frac{2\kappa(T_{g1} + T_0)}{\delta_0(\delta_0 + \delta_1)} \qquad \ldots \ldots \ldots \ldots \ldots (15)$$

where $\kappa = \lambda / C$. The same general technique is used to provide the corresponding predictive equations for the temperatures of the other soil layers.

In the three-layer soil model illustrated in Fig. 1, all three soil temperatures are treated as prognostic variables during the integration of the model, with the condition that the soil heat flux is zero at the lower boundary ($G_3 = 0$). An alternative popular lower-boundary condition is to hold the temperature of the bottom layer at its initialized value. This approach is used, for example, in the four-layer model in the current Meteorological Office fine-mesh model (Carson 1986) and also in the three-layer soil model used at the European Centre for Medium Range Weather Forecasts (ECMWF) (Blondin 1986).

The selection of 'representative' soil thermal characteristics (C and λ, and hence κ) and suitable soil-layer depths ($\delta_0, \delta_1 \ldots \delta_{n-1}$ where n is the number of explicitly resolved layers in the soil) remains difficult, empirical and highly subjective. On the basis of a comprehensive study of the amplitude and phase responses of multi-layer soil schemes to periodic surface temperature forcing, Warrilow *et al.* (1986) have recommended a four-layer soil temperature scheme. Their paper gives full details of how the appropriate soil parameters were selected. The values which have been chosen for the fourth-annual cycle version of the Meteorological Office 11-layer AGCM are:

$$C = 2.34 \times 10^6\ \mathrm{J\,m^{-3}\,K^{-1}},\ \lambda = 0.56\ \mathrm{W\,m^{-1}\,K^{-1}},\ \kappa = \lambda / C = 2.39 \times 10^{-7}\ \mathrm{m^2\,s^{-1}},$$

$$\delta_0 = 0.037\ \mathrm{m},\ \delta_1 = 0.143\ \mathrm{m},\ \delta_2 = 0.516\ \mathrm{m},\ \delta_3 = 1.639\ \mathrm{m}$$

where $\delta_0 = (\kappa P / \pi)^{1/2}$ with $P = 0.2048$ days (4.8 hours).

Further information about values used in specific models can be found in, for example, Blondin (1986) and Carson (1982, 1986).

2.4 The land-surface temperature

It is usually assumed that the land-surface temperature is a well-defined and unique property of any natural land surface, and that the same T_0 is appropriate as:

(a) the radiative surface temperature used to compute the thermal emission from the earth's surface (see section 3.2 of Part I);

(b) the surface temperature as used in the extrapolated atmospheric boundary-layer profiles and surface-flux formulae (see section 4.2 of Part I);

(c) the surface soil temperature which is involved in the computation of the soil heat flux (see previous sections of this article).

The 'surface temperatures' implied by these different physical processes at the surface must be closely related but they are not necessarily the same. The ambiguity and difficulty in defining surface temperature become even greater when the surface has a vegetative canopy. Suffice it to state that at present the problem is very poorly understood and that many observational and theoretical studies are needed before any significant differences between the 'T_0' can be clearly delineated and incorporated sensibly in AGCMs and NWPMs.

3. Surface hydrology and the soil water budget

Most of the current generation of AGCMs and NWPMs now include some form of 'interactive' surface hydrology, usually of a very rudimentary nature. Such parametrizations are termed 'interactive' in the sense that the soil has some recognized hydrological property that is allowed to vary in response to the model's continuously evolving atmospheric state and surface boundary conditions and which in turn exerts both direct and indirect influences on the surface fluxes themselves. The most common practice is to define a variable 'soil moisture content' for some notional depth of surface soil layer which is constrained at all times to satisfy the surface moisture flux balance expressed as

$$M_0 = P_r - E - Y_0 \qquad \ldots \ldots \ldots \ldots \ldots \ldots (16)$$

where P_r is the intensity of surface rainfall, E is the surface evaporation rate (turbulent flux of water vapour), Y_0 denotes intensity of surface run-off and M_0 represents the net mass flux of water into the soil layer.

As in the case of the surface radiative fluxes, $R_S{\downarrow}$ and $R_L{\downarrow}$, in the context of the surface energy balance (discussed in section 3 of Part I), P_r is regarded here as an externally determined component of the surface moisture balance (equation (16)). Accurate evaluation of P_r is, of course, of crucial importance in establishing a realistic surface moisture balance and also, through the coupling discussed above, a realistic surface energy balance. The other processes involved in the hydrology of a bare soil, including evaporation, surface run-off, and transport and storage of water in the soil, are generally very complex and not so well understood nor as simple to parametrize sensibly as the individual terms in the surface energy balance. The very-small-scale spatial inhomogeneities within a typical soil layer appear to be more important in the determination of soil moisture movement than for the heat flow, and this presents formidable difficulties when trying to formulate a parametrization based soundly on underlying physical and dynamical hydrological principles. This is particularly so when one-dimensional hydrological models are applied to catchment-sized or typical AGCM/NWPM grid-box areas. Hence the importance of the HAPEX–MOBILHY project (André *et al.* 1986) aimed at studying the hydrological budget and evaporation flux at the scale of an AGCM grid square, i.e. $10^4\ \mathrm{km}^2$. A 2½-month special observing period should provide detailed measurements of the relevant atmospheric fluxes and intensive remote sensing of surface properties. The main objective of the programme is to provide a data base against which parametrizations of the land-surface water budget can be developed and tested.

A proper discussion of the surface and subsurface hydrology of natural soils is beyond the scope of this paper. For this the reader is referred to the recent fuller expositions by, for example, Brutsaert (1982a, b), Dooge (1982), Eagleson (1982) and Dickinson (1984) in which the problems of areal representation of hydrological processes are specifically discussed. The remainder of this section is restricted to an introduction to the most simple form of the basic equations which govern the movement of water in the soil, and brief descriptions of some specific formulations for soil water transport, evaporation, and surface run-off. These examples, although chosen quite subjectively, should nevertheless give an indication of the general tenor and level of many of the current attempts to parametrize subgrid-scale hydrological processes.

3.1 *Water transport in a homogeneous soil*

There are various interrelated measures of soil moisture content, two of which are:

(a) the soil moisture concentration, χ, defined as the mass of water per unit volume of soil (units kg m^{-3}), and

(b) the volumetric soil moisture concentration, χ_v, defined as the volume of water per unit volume of soil (a dimensionless quantity).

Therefore $\chi = \rho_W \chi_v$ where ρ_W is the density of water. These are very appropriate measures in parametrizations based on simulating changes in the water mass of a specified layer of soil.

In an analogous fashion to the treatment of soil heat conduction discussed earlier, it is convenient to consider a grossly simplified hydrology of a spatially homogeneous soil layer with no horizontal water movement and no melting or freezing within it. This restricted and idealized one-dimensional problem is governed by the equation of continuity and the flux–gradient relation (Darcy's law). Combining these two equations yields Richards's equation for the vertical movement of water in an unsaturated soil. Solving the full Richards's equation for reasonable boundary conditions is by no means easy. However, proposals do exist which simplify the problem in a highly empirical way which is difficult to justify in all but the most idealized circumstances. In such cases Richards's equation takes the form of a diffusion equation for soil water

$$\frac{\partial \chi_v}{\partial t} = \frac{\partial}{\partial z_g}\left(\kappa_W \frac{\partial \chi_v}{\partial z_g}\right) - \frac{\partial K}{\partial z_g}$$

where K is the hydraulic conductivity of soil (units m s^{-1}) and κ_W is a moisture diffusivity of soil (units m^2 s^{-1}).

Prognostic equations for soil moisture content based on an idealized hydrology are beginning to appear in AGCMs and NWPMs (e.g. Dickinson 1984 and Warrilow *et al.* 1986). One particular multi-layer soil hydrology scheme which has attracted considerable support from numerical modellers is the force-restore treatment of Deardorff (1978) in which he postulates equations for soil moisture transport of a form directly analagous to the corresponding force-restore equations for soil temperature (see equation (12)). An effective three-layer version of Deardorff's approach is used, for example, in the ECMWF model (Blondin 1986). However, the most common current approach to modelling soil moisture content is probably still that based on a single surface-soil layer and a more detailed discussion of that example will suffice here.

3.2 *Single-layer soil hydrology models*

A common rudimentary approach to the parametrization of the hydrological processes at a bare-soil surface is to monitor the change of soil moisture content in a single shallow surface-soil layer of notional depth δ_W as depicted schematically in Fig. 2.

Let m_W denote the mass of liquid water per unit lateral area in the soil layer of depth δ_W, that is

$$m_W = \int_0^{\delta_W} \chi \mathrm{d}z_g = \hat{\chi}\delta_W = \rho_W \hat{\chi}_v \delta_W = \rho_W d_W \qquad \ldots \ldots \ldots (17)$$

where this equation also defines a layer-mean soil moisture concentration, $\hat{\chi}$, a layer-mean volumetric soil moisture concentration, $\hat{\chi}_v$, and a representative depth of water in the layer, d_W.

The continuity equation for χ is

$$\frac{\partial \chi}{\partial t} = -\frac{\partial M}{\partial z_g}$$

where M is the vertical mass flux of water. Integrating over the depth δ_W gives the water mass balance equation for the surface layer

$$\frac{\partial m_W}{\partial t} = M_0 - M_1$$

where M_0 is the net mass flux of water into the soil layer and M_1 is the vertical mass flux of water at the base of the surface layer. Substituting for M_0 from equation (16) yields

$$\frac{\partial m_W}{\partial t} = P_r - E - Y_0 - M_1 \qquad \ldots \ldots \ldots \ldots \ldots (18)$$

It should be noted that, in deriving equation (18), all horizontal fluxes of soil water have been neglected apart from the surface run-off, Y_0.

With the terms E, Y_0 and M_1 evaluated according to a particular model's approach selected from the wide range of methods available, and P_r determined by some other parametrization in the model, equation (18) can be solved either in a simple explicit fashion or by more subtle implicit methods to determine the change in m_W (and hence $\hat{\chi}_v$, $\hat{\chi}$, d_W, etc.).

With P_r regarded in the present context as determined externally, then it remains here to illustrate, with the aid of specific examples, some of the problems of formulating parametrizations for E, Y_0 and M_1.

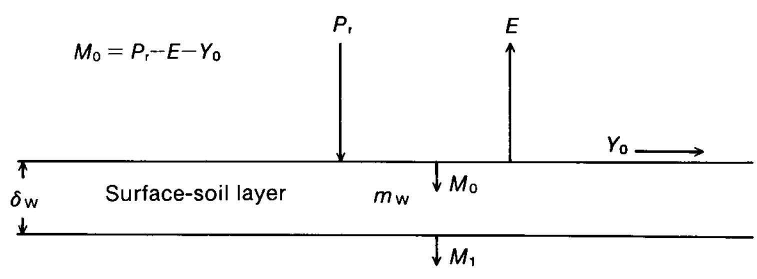

Figure 2. Schematic representation of moisture balance in a single shallow surface-soil layer. For explanation of symbols see text.

3.3 *Evaporation at a bare-soil surface, E*

In principle, E can be obtained as the residual flux from either the surface energy balance (equation (1)) or the surface moisture balance (equation (16)) and there are many empirical formulae for estimating E based on such approaches. A very useful introduction to the large variety of methods available can be found, for example, in Rosenberg *et al.* (1983); for more detailed discussions see, for example, Eagleson (1982) and Brutsaert (1982a, b). However, in this introduction to interactive soil temperature and soil moisture content parametrizations in AGCMs and NWPMs, I have selected the soil flux terms G_0 and M_0 as the residual components in the surface balance equations (1) and (16), and assumed implicitly that E can be evaluated in some independent manner.

Indeed, the method of estimating E has already been considered in Part I where E was ultimately parametrized in the bulk-aerodynamic form of

$$E = -\rho C_E V(z_l)\{q(z_l) - q_0\} \qquad \ldots \ldots \ldots \ldots (19)$$

with the recommendation that the bulk transfer coefficient, C_E, be evaluated from the Monin–Obukhov similarity theory (see section 4.2 of Part I). It was, however, also noted that the surface value of the specific humidity, q_0, required explicitly in equation (19) (and also, for example, in determining the bulk Richardson number and hence C_E) is not easy to determine. To overcome this problem it is standard practice to imply a value of q_0 through relations with $q_{sat}(T_0)$, the saturation specific humidity at the surface.

Two common methods are:

(a) to specify a surface relative humidity, r_0, such that $q_0 = r_0\, q_{sat}(T_0)$;

(b) to evaluate a potential evaporation rate $E_p = -\rho C_E V(z_l)\{q(z_l) - q_{sat}(T_0)\}$ and to specify an empirical 'moisture availability function', β, (usually ranging from 0 for an arid surface to 1 for a saturated surface) such that the actual evaporation rate is given by $E = \beta E_p$.

The second method is by far the more commonly adopted. For a discussion and comparison of the two approaches see, for example, Nappo (1975) and for examples of their use in specific AGCMs see Carson (1982). It is worth noting in passing that an alternative relation in the spirit of method (b) is used for computational convenience in some models and that is $\Delta q(z_l) = q(z_l) - q_0 = \beta\{q(z_l) - q_{sat}(T_0)\}$ which implies that $q_0 = \beta q_{sat}(T_0) + (1-\beta)q(z_l)$. The reasons for preferring this formulation are discussed in Carson (1982) (and more fully in Carson and Roberts 1977).

The most common method now employed is to express β as a simple linear function of the soil moisture content in the surface layer, $\hat{\chi}_v$, with $\beta = 1$ once $\hat{\chi}_v$ reaches a critical value, $\hat{\chi}_{v,c}$ (see Fig. 3(a)). However, Warrilow *et al.* (1986) have suggested an alternative formulation which involves the notion of a wilting point, $\hat{\chi}_{v,w}$ (see Fig. 3(b)). This represents a level of soil moisture concentration below which further water cannot be removed from the soil by the normal processes.

The critical value $\hat{\chi}_{v,c}$ is not well defined but for simplicity, and in line with previous practice (e.g. Carson 1982), it is given by

$$\hat{\chi}_{v,c} = \hat{\chi}_{v,w} + \tfrac{1}{3}(\hat{\chi}_{v,f} - \hat{\chi}_{v,w})$$

where $\hat{\chi}_{v,f}$ is a nominal field capacity used to define $\hat{\chi}_{v,c}$. The particular values of $\hat{\chi}_{v,w}$, $\hat{\chi}_{v,c}$ and $\hat{\chi}_{v,f}$ being used globally in the Meteorological Office AGCM, which assumes, for hydrological purposes only, a single surface layer of soil and nominal depth $\delta_W = 1$ m, are:

$$\hat{\chi}_{v,w} = 0.08, \qquad \hat{\chi}_{v,c} = 0.13, \qquad \hat{\chi}_{v,f} = 0.23.$$

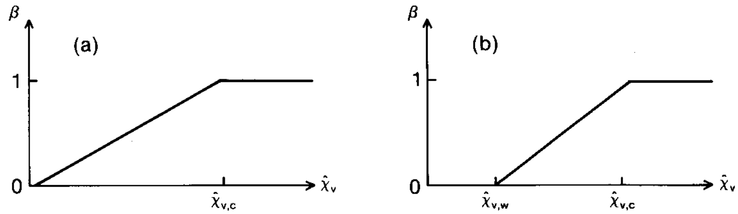

Figure 3. Variation of moisture availability function, β, with soil moisture content, $\hat{\chi}_v$, for (a) the most common scheme and (b) the scheme used in the Meteorological Office 11-layer model (see Warrilow *et al.* 1986). For explanation of other symbols see text.

Although β has been expressed in terms of $\hat{\chi}_v$, there is no reason why it should not be simply reformulated in terms of any of the other measures of soil moisture content, the most common of which is d_W.

It should be borne in mind that the more complex and real practical issue to be addressed is that of determining the actual evapotranspiration from partially vegetated surfaces and not simply the evaporation from a bare-soil surface.

3.4 *Surface run-off,* Y_0

Surface run-off is yet another of the complex surface hydrological processes which is treated very simplistically in current AGCMs and NWPMs. In models employing the single-layer water mass balance (equation (18)), the simplest approach is the so-called 'bucket method' for run-off in which rainfall (modified by evaporation loss) is allowed to increase the soil moisture content until $\hat{\chi}_{v,f}$ is reached. Any further attempt to increase $\hat{\chi}_v$ beyond the field capacity is implicitly assumed to be run-off water (a combination of Y_0 and M_1) which plays no further part in the model's hydrological cycle. This identifies the original role played in these simple hydrological parametrizations by the field-capacity term, in addition to its use to define $\hat{\chi}_{v,c}$. For a selection of the crude and highly empirical formulations used in specific AGCMs see Carson (1982).

A novel, but still relatively simple, parametrization has been developed by Warrilow *et al.* (1986) for use in the Meteorological Office AGCM. It is based, with considerable simplification, on a scheme proposed by Milly and Eagleson (1982). An attempt has been made to allow for the spatial variability of rainfall, since use of grid-box averages would give a marked underestimation of the surface run-off. It is assumed that rain only falls on a proportion, μ, of the grid area, where μ is chosen arbitrarily as 1 for the model's 'large-scale dynamic rain' and 0.3 for its 'convective rain'. The local rainfall rate, P_{rl}, throughout the grid-area is treated statistically by assuming that it has a probability density function given by

$$\mathrm{f}(P_{rl}) = \frac{\mu}{P_r} \exp\left(-\frac{\mu P_{rl}}{P_r}\right)$$

where P_r is the model's grid-point rainfall rate which is taken to represent the average grid-area rainfall. Whenever P_{rl} is greater than the surface infiltration rate, F, the local surface run-off is $Y_{0l} = P_{rl} - F$. Integration of Y_{0l} over all values of P_{rl} yields an expression for the total run-off rate for a grid area $Y_0 = P_r \exp(-\mu F/P_r)$ where Warrilow *et al.* (1986) have taken F to have a global value such that $F/\rho_W = 13.0\ \mathrm{mm\ h^{-1}}$.

3.5 *The vertical mass flux of water at the base of the surface layer,* M_1

The simplest single-layer approaches typically assume explicitly that in equation (18) M_1 is negligible, or implicitly that it is combined with Y_0 to give a total run-off. In the scheme of Warrilow *et al.* (1986), adapted from Milly and Eagleson (1982), M_1, referred to as the gravitational drainage from the base of the surface layer, is acknowledged as a separate hydrological component in equation (18) that has to be parametrized.

Warrilow *et al.* (1986) have argued, somewhat speculatively, that for horizontal averaging over a typical AGCM grid area M_1 can be represented simply by $M_1 = \rho_W\{K(\chi_v)\}_{z_g = \delta_W}$ with the further assumption that χ_v is effectively spatially homogeneous in the surface-soil layer so that $M_1 = \rho_W K(\hat{\chi}_v)$. Their particular prescription of the hydraulic conductivity as a function of $\hat{\chi}_v$, attributed to Eagleson (1978), is

$$K(\hat{\chi}_v) = K_{sat}\left(\frac{\hat{\chi}_v - \hat{\chi}_{v,w}}{\hat{\chi}_{v,sat} - \hat{\chi}_{v,w}}\right)^c$$

where $\hat{\chi}_{v,sat}$ is termed the saturation value of $\hat{\chi}_v$, K_{sat} is the saturation conductivity (i.e. $K(\hat{\chi}_{v,sat})$) and c is an empirically derived constant. The particular values adopted globally by Warrilow *et al.* (1986) are:

$$\hat{\chi}_{v,sat} = 0.445, \qquad K_{sat} = 13.0 \text{ mm h}^{-1}, \qquad c = 6.6.$$

4. Snow-covered surfaces

A particular class of non-vegetated land surfaces which have their own very special characteristics and exercise significant influence on the climate system over a wide range of time-scales is that comprising snow-covered (and ice-covered) surfaces. As in the case of land-surface hydrology, it is generally true that little attention has yet been given to the representation in AGCMs and NWPMs of the special physical processes associated with such surfaces. However, I am confident that this particular area of the wider problem will receive increasing attention in the near future.

According to Kuhn (1982), in the course of the year about 50% of the earth's land surface is covered by snow or ice. He also comments that, although the polar ice sheets contain about 99% of the earth's freshwater ice by mass, nevertheless the seasonal snow cover with its large areal extent and its high spatial and temporal variability may have an equal or even greater impact on the atmospheric circulation. Undoubtedly then, a key issue will be how to deal sensibly with partial and rapidly changing snow cover, particularly in complex terrain, over the area of a typical grid box in a large-scale numerical model. The proper treatment of the processes associated with snow-covered surfaces is a major topic in its own right. The brief comments here are no more than a postscript to the main discussion of bare-soil surfaces. For fuller expositions of the varied and complex characteristics and the effects of snow and its associated physical processes see, for example, Martinelli (1979), Male (1980), Gray and Male (1981) and International Glaciological Society (1985). For discussions of snow-covered surfaces aimed specifically at the AGCM parametrization problem see, in particular, Kuhn (1982) and Kotliakov and Krenke (1982).

4.1 *Special conditions at snow- and ice-covered surfaces*

Kuhn (1982) has listed the special conditions for snow and ice layers as:

(a) the surface temperature cannot exceed the melting temperature of ice;
(b) evaporation and sublimation take place at the potential rate;
(c) the short-wave albedo is generally high;
(d) the medium is permeable to air and water and transparent to visible radiation;
(e) the snow pack is a good thermal insulator;
(f) the layer has a high storage capacity for heat and water;

(g) the roughness of the surface is extremely low;
(h) generally, the atmospheric surface layer over snow or ice is stably stratified.

Note that conditions (a)–(h) impinge on every aspect of the parametrization problem already discussed. The remainder of this section retraces the previous route and indicates briefly where modifications to the parametrizations are typically introduced into AGCMs and NWPMs in recognition of snow (or ice) covering the surface. In general, the thermal and hydrological properties of the snow pack are represented very simply and crudely in such models.

4.2 *The physical properties of snow- and ice-covered surfaces*

(a) *Short-wave albedo, α.* It is firmly established that the physical coupling between snow and ice cover, albedo, and the surface temperature is one of the most important feedback mechanisms to include in an AGCM. An important characteristic of snow- and ice-covered surfaces is their high reflectivity compared with other natural surfaces such that even a thin covering of fresh snow can alter significantly the albedo of a landscape. The local albedo of a snow-covered surface is very variable and is a complicated function of many factors including the age of the snow pack (α decreases markedly as the snow becomes compacted and soiled), the wavelength and angle of the incident radiation, and even diurnal cycles in the state of the snow surface, particularly when conditions are right for surface melting. The albedo may lie anywhere in the range from 0.95 for freshly fallen snow to about 0.35 for old, slushy snow (see, for example, Kondratyev *et al.* 1982).

At present the coupling between snow and ice and the surface albedo is generally prescribed very simply. Three types of snow- or ice-covered surfaces are generally acknowledged: surfaces with instantaneously variable depth of snow either predicted or implied, permanent or seasonally prescribed snow- and ice-covered land surfaces, and permanent or seasonally prescribed areas of sea-ice. The last category is not the concern of this paper. For models that 'carry' a snow depth a common approach is that of Holloway and Manabe (1971) who, following Kung *et al.* (1964), used:

$$\alpha = \begin{cases} \alpha_L + (\alpha_S - \alpha_L)\, d_{SW}^{1/2} & d_{SW} < 1\ \text{cm} \\ \alpha_S & d_{SW} \geq 1\ \text{cm} \end{cases}$$

where α_L is the snow-free land-surface albedo, α_S is the albedo (assumed in this case to be 0.60) of a deep-snow surface and d_{SW} is the water equivalent depth of snow (here expressed in centimetres). No allowance is usually made for the varying density of a snow pack, and d_{SW} is assumed typically to be about 1/10 of the actual snow depth.

In some models which predict and monitor snowfall, a single albedo value is used for any non-zero depth of snow (see Carson 1982, 1986). The first snowfall on a previously snow-free surface results in an immediate increase in surface albedo which will tend, at least initially, to accelerate the positive feedback of a further lowering of the surface temperature with an enhanced probability of further snow accumulation. Typical model values for land- and sea-ice are in the range 0.5–0.8 (see Carson 1982, 1986).

(b) *Long-wave emissivity, ϵ.* Kuhn (1982) states that this can be assumed to be unity for all practical purposes.

(c) *Surface roughness length, z_0.* The effective z_0 for extensive, uniformly covered snow and ice fields and the 'local' value of z_0 for snow-covered, simple heterogeneous terrain may indeed be very small (of the order of 10^{-3} m or less). However, in general, the effective areal z_0 of natural, heterogeneous and complex terrain with varied relief and vegetation is very difficult to determine and may be affected greatly or insignificantly by different degrees of snow cover. There is little scope for useful discussion of this problem in a global, large-scale modelling context except to note that, in principle, snow and ice cover can alter z_0.

(d) *Thermal properties of snow.* As noted above, a snow pack is generally a good thermal insulator for the soil below, but to capture this effect in a climate model implies a delineation and explicit modelling of the heat conduction (and the hydrology) in and between the two media. In general, the thermal and hydrological properties of snow and ice layers are treated very simply, if at all, in AGCMs and NWPMs. The thermal properties of a snow pack will, like its density and albedo, depend in a complicated fashion on many factors. Values thought to be appropriate for snow are given in Table I for comparison with the range of soil values also included there.

4.3 *Surface energy and mass flux balances at a snow-covered surface*

The surface energy flux balance (equation (1)) is modified for complete snow cover such that

$$G_0 = R_N - H - Q_s - Q_f \qquad \text{(20)}$$

where $Q_f = L_f S$ represents the latent-heat flux required to affect phase changes associated with melting or freezing at the surface (where S is the rate of snow melt or ice melt and L_f is the latent heat of fusion), $Q_s = L_s E_s$ represents the latent-heat flux due to surface sublimation by turbulent transfer (where E_s is the latent heat of sublimation ($L_s = L_e + L_f$)), and G_0 is now strictly the flux of heat into the snow layer at its upper surface.

A simple budget equation, corresponding to that used for soil moisture content in a single-soil layer (equation (18)), is also used for snow on the 'surface', viz.

$$\frac{\partial m_S}{\partial t} = P_S - E_s - S \qquad \text{(21)}$$

where P_S is the intensity of snowfall at the surface and m_S is the mass of snow lying per unit area of the surface. Therefore m_S is treated like m_W as a surface prognostic variable and is often represented as a snow depth, d_S, or more commonly as an equivalent depth of water, d_{SW}, such that $m_S = \rho_S d_S = \rho_W d_{SW}$ where ρ_S is the density of snow. Although it is recognized that the density of a snow pack varies, this again is a complicated issue in its own right and it is quite common practice in large-scale numerical models to assume simply that $\rho_S \simeq 0.1\ \rho_W$.

Equation (21) is usually complemented by the surface-layer balance equation for the soil moisture content (equation (18)) modified to include the snow-melt term, i.e.

$$\frac{\partial m_W}{\partial t} = P_r - E + S - Y_0 - M_1 . \qquad \text{(22)}$$

Each model has its own system of checks and algorithms for deciding which of the terms in equations (21) and (22) are in force simultaneously. One popular approach is as follows. When snow is lying, T_0 is not allowed to rise above 273 K and the snow depth accumulates without limit or decreases according to the net value of ($P_S - E_s$). If, however, snow is lying and the solution of the heat balance equation, equation (20), excluding the term Q_f, produces an interim surface temperature value $T_0' > 273$ K then sufficient snow (if available) is allowed to melt to maintain $T_0 = 273$ K. The heat required to melt the snow and reduce T_0 to 273 K can be evaluated by specifying an effective surface thermal capacity of the snow pack (cf. equation (5)) such that

$$Q_f = L_f S = C_{\text{eff, S}} \left(\frac{T_0' - 273}{\Delta t} \right) \qquad \text{(23)}$$

where Δt is the appropriate model time step. The change in the water equivalent snow depth, Δd_{SW}, resulting from the melting is determined from equations (22) and (23) such that

$$\Delta m_S = \rho_W \Delta d_{SW} = -S\,\Delta t$$
$$= -\frac{C_{eff,\,S}}{L_f}(T_0' - 273).$$

It is usually assumed that the snow pack has no moisture-holding capacity; all melted snow is added directly to the soil moisture content (through equation (22)) following the corresponding reduction, Δd_{SW}, in the snow depth. In all cases it is only when the snow disappears through melting or sublimation that evaporation of moisture is allowed to resume at the surface.

5. Concluding remarks

It should be evident that, in many respects, the representations of land-surface processes in AGCMs and NWPMs are still rather crude and simple. The demands for improvements will come from both climate modelling studies and numerical weather forecasting. Indeed, the steadily increasing number of studies with AGCMs has already amply demonstrated the sensitivity of such models to surface properties and processes (see, for example, recent reviews by Mintz (1984), Rowntree (1983, 1984) and Rowntree *et al.* (1985)). Parametrizations thought adequate at present will undoubtedly be seen to be deficient in models which couple interactively further components of the climate system. This is already apparent with respect to sea–air interactions in coupled ocean–atmosphere models. Although the major developments in the longer term are more likely to come from climate modelling studies, nevertheless valuable feedback is being obtained from the continuous close scrutiny of the various models' performances in the acutely critical arena of operational weather forecasting — especially of local, near-surface variables such as wind and temperature.

A schematic résumé of the processes, variables and parameters introduced in this discussion of the specification of parametrizations for simple, non-vegetated land surfaces is given in Table II.

Table II. *Schematic résumé of the processes, variables and parameters involved in the specification of parametrizations at simple, non-vegetated land surfaces. For explanation of symbols see text.*

	Radiative	Thermal	Hydrological	Dynamical
'External forcing'	$R_{S\downarrow}$, $R_{L\downarrow}$		P_r, P_S	
Atmospheric variables		$\theta(T)$	q	$\underline{V}$
Surface variables	T_0	T_0	q_0, d_S	
Surface parameters	α, ϵ	z_0	z_0, β	z_0
Surface fluxes	R_N	H $Q = L_eE$ $Q_f = L_fS$ $Q_s = L_sE_s$ G_0	 E, E_S S M_0, Y_0	$\underline{\tau}$
Subsurface fluxes		G	M	
Subsurface parameters		λ, C	δ K Significant values of χ_v	
Subsurface variables		T_g	χ_v	

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Wind measurements from the Earth Resources Satellite (ERS-1)

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Summary

Within 3 years, a new form of satellite data will become accessible to operational meteorology; near-surface wind vectors. ERS-1 will have a complement of microwave instruments which will provide all-weather coverage of the world's oceans. The Meteorological Office has been involved in the design of this experiment and is currently assisting the European Space Agency with the development of data-processing algorithms. After launch the data will be available, within a few hours of measurement, to be incorporated into numerical models.

1. Introduction

The European Space Agency (ESA) is planning to launch its first remote sensing satellite, the Earth Resources Satellite (ERS-1), early in 1990. Its main complement of instruments are various types of radar, operating at microwave frequencies (5–15 GHz), plus a passive infra-red instrument, the Along

Track Scanning Radiometer (ATSR), which is being built by the Rutherford Appleton Laboratory with Meteorological Office assistance. Table I summarizes the instruments to be carried, their main purpose and expected accuracies. Primarily, the satellite's mission is to sense several oceanographic parameters, both for experimental purposes and as a demonstration that such instruments could have a role to play in operational meteorology and oceanography on the global scale.

Of these instruments, the one most applicable to meteorology is the Advanced Microwave Instrument (AMI) package in its wind mode of operation. This is designed to sense the near-surface (equivalent 10 m) wind vector over the oceans. This article will describe the principles of operation of this type of sensor, and the processing necessary to interpret its raw measurements into familiar geophysical units. Any form of instrument needs calibrating, and the AMI is no exception, so the plans to achieve this will also be outlined. Finally, the use to be made of the resulting data within the Meteorological Office will be mentioned.

Table I. *ERS-1 main geophysical measurement objectives*

Main geophysical parameter	Range	Accuracy	Main instrument
Wind field			
Velocity	4–24 m s^{-1}	± 2 m s^{-1} or 10% whichever is greater	AMI wind scatterometer and altimeter
Direction	0–360°	± 20°	AMI wind scatterometer
Wave field			
Significant wave height	1–20 m	± 0.5 m or 10% whichever is greater	Altimeter
Wave direction	0–360°	± 15°	AMI wave mode
Wavelength	50–1000 m	20%	AMI wave mode
Earth surface imaging			
Land/ice/coastal zones, etc.	80 km (minimum swath width)	Geometric/radiometric resolutions 30 m/2.5 dB	AMI Synthetic Aperture Radar (AMI-SAR)
Altitude over ocean	745–825 km	2 m absolute ± 10 cm relative	Altimeter
Satellite range		± 10 cm	Precise Range and Range-Rate Experiment (PRARE)
Sea surface temperature	500 km swath	± 0.5 K	Along Track Scanning Radiometer (Infra-Red) (ATSR/IR)
Water vapour	25 km spot	10%	Microwave sounder (part of ATSR experiment) (ATSR/M)

2. Theoretical and experimental background

As soon as microwave radar became widely used in the 1940s, it was found that at low elevation angles surrounding terrain (or at sea, waves) caused large, unwanted echoes. Ever since, designers and users of radar equipment have sought to reduce this noise (Harrold 1974). Researchers investigating the effect found that the back-scattered echo became larger with increasing wind speed, so opening the possibility of remotely measuring the wind (Krishen 1971, Jones and Schroeder 1978). Radars designed to measure this type of echo are known as 'scatterometers'.

The back-scattering is due principally to in-phase reflections from a rough surface; at microwave frequencies this arises from the small ripples (cat's paws) generated by the instantaneous surface wind stress. The level of back-scatter from an extended target, such as the sea surface, is generally termed the Normalized Radar Cross-Section (NRCS), or σ°. For a given geometry and transmitted power, σ° is proportional to the power received back at the radar. Experimental evidence from scatterometers operating over the ocean shows that σ° increases with surface wind speed (as measured by ships or buoys), decreases with incidence angle, and is also dependent on the radar beam angle relative to wind direction. Fig. 1 is a plot of σ° aircraft data against wind direction for various wind speeds. Direction 0° corresponds to looking upwind, 90° to crosswind and 180° to downwind.

Over the past three winters, ESA has co-ordinated a number of experiments to confirm these types of curves at 5.3 GHz, the AMI operating frequency. Several aircraft scatterometers have been flown close to instrumented ships and buoys, in the North Sea, the Atlantic and the Mediterranean. The σ° data are then correlated with the surface winds, which have been adjusted to a common anemometer height of 10 m (assuming neutral stability). An empirical model function has been fitted to these data (Long 1986) of the form

$$\sigma^\circ = a_0 U^{\gamma} (1 + a_1\cos\phi + a_2\cos2\phi) \qquad \ldots \ldots \ldots \ldots \ldots (1)$$

where the coefficients a_0, a_1, a_2 and γ are dependent on the incidence angle. This model relates the neutral stability wind speed at 10 m, U, and the wind direction relative to the radar, ϕ, to the NRCS.

It may also be the case that σ° is a function of sea surface temperature, sea state and surface slicks (natural or man-made), but these parameters have yet to be demonstrated as having any significant effect on the accuracy of wind vector retrieval.

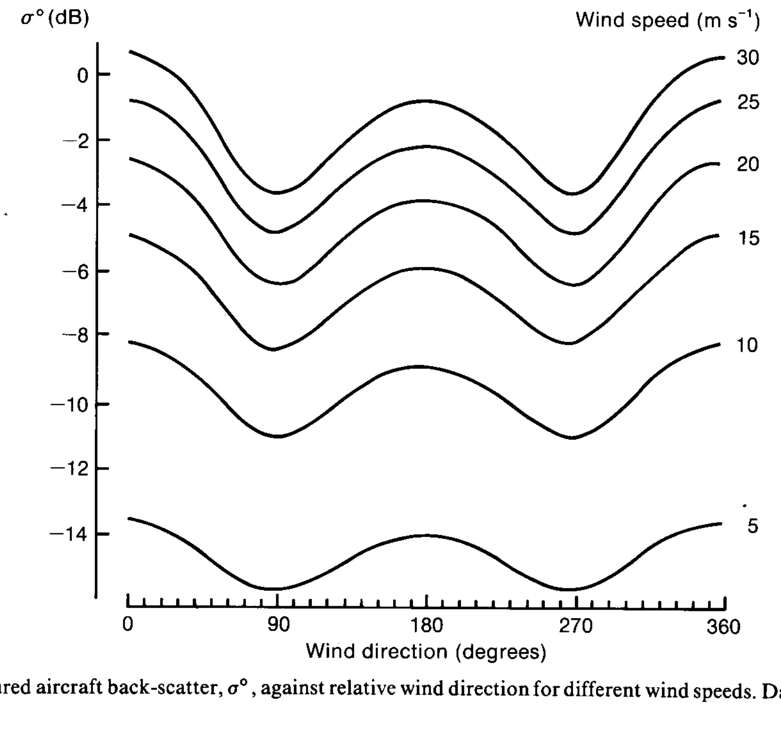

Figure 1. Measured aircraft back-scatter, σ°, against relative wind direction for different wind speeds. Data for 13 GHz, vertical polarization.

3. The AMI wind scatterometer

Since σ° shows a clear relationship with wind speed and direction, in principle, measuring σ° at two or more different azimuth angles allows both wind speed and direction to be retrieved. The first wind scatterometer to be flown on a satellite — SEASAT — was in 1978 and ably demonstrated the accuracy of this new form of measurement (Offiler 1984). The SEASAT-A Satellite Scatterometer (SASS) instrument used two beams either side of the spacecraft; the ERS-l AMI scatterometer uses a third, central beam to improve wind direction discrimination, but is only a single-sided instrument, so its coverage is less. Fig. 2 shows the geometry of the ERS-1 scatterometer.

The three antennae each produce a narrow beam of radar energy in the horizontal, but wide in the vertical, resulting in a narrow band of illumination of the sea surface across the 500 km width of the swath. As the satellite travels forward, the mid beam, then the aft beam will measure from the same part of the ocean as the forward beam. Hence each part of the swath, divided into 50 km squares, will have three σ° measurements taken at different relative directions to the local surface-wind vector.

Fig. 3 shows the coverage to be expected of the scatterometer for the North Atlantic over 24 hours. These swaths are not static, but 'move' westwards to fill in the large gaps on subsequent days. Even so, the coverage will not be complete, owing to the relatively small swath width in relation to, say, the Advanced Very High Resolution Radiometer (AVHRR) imager on the NOAA satellites. However, there will potentially be a wind available every 50 km within the coverage area, globally, and ESA is committed to delivering this to operational users within three hours of measurement time. The raw instrument data are to be recorded on board and replayed to ESA ground stations each orbit, the principal station being at Kiruna in northern Sweden where the wind vectors will be derived.

4. Wind vector retrieval

As already mentioned, the AMI scatterometer principally measures the level of back-scatter at a given location at different azimuth angles. Since we know the geometry, such as range and incidence angles, it ought to be possible to use the model function (equation (1)) to extract the two pieces of information required — wind speed and direction — using appropriate simultaneous equations. However, in practice this is not feasible; the three values of σ° will have a finite-measurement error, and the function itself is highly non-linear. Indeed the model, initially based on aircraft data, may not be applicable to all circumstances.

The wind speed and direction must be extracted numerically, usually by comparing the measured values of σ° to those from the model function, using an initial estimate of wind speed and direction, then refining the estimate so as to minimize the σ° differences. Starting at different first-guess wind directions, the numerical 'solution' can converge on up to four distinct, or ambiguous, wind vectors, although often there are only two obviously different ones — usually about 180° apart. One of these is the 'correct' solution, in that it is the closest to the true wind direction and within the required root-mean-square accuracies of 2 m s^{-1} and 20°.

Selecting which of the ambiguous set of solutions is the correct one, when the true wind is unknown (as is the case for an operational system), is termed ambiguity removal. This is merely assigning a probability of correctness to each solution and choosing the one with the highest probability. This may sound easy, but to do this with a consistent high skill (ideally one would want to choose correctly on 100% of cases) is a difficult task. A first stage is to see which extracted vector best fitted the model function, ranking them in order of probability, and choosing the first rank. Simulations (Offiler 1985) have indicated that this technique might be expected to have a skill rating of not more than 70%. (Compare this with SASS which, with only two beams, could only manage a random 25% skill.)

Further processing may be done by objectively seeking areas of the swath where the first rank shows some consistency in direction, and selecting a different rank for nearby cells which do not at first agree with this general trend. This may yield a skill score of 90% or so in most cases, but areas which have been consistently ranked wrongly to start with will only be reinforced, so drastically reducing the skill in picking the right solution.

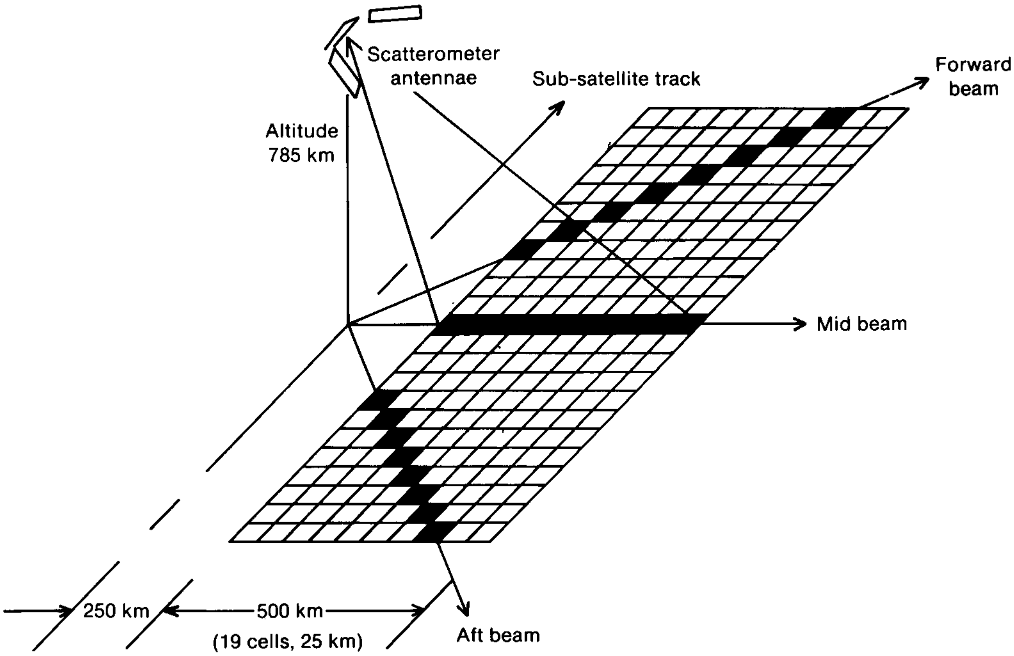

Figure 2. ERS-1 wind scatterometer geometry. For clarity the 500 km wide swath is shown as 10 non-overlapping cells. In reality there are 19 cells across the swath, on a 25 km grid, each one being 50 km × 50 km in size.

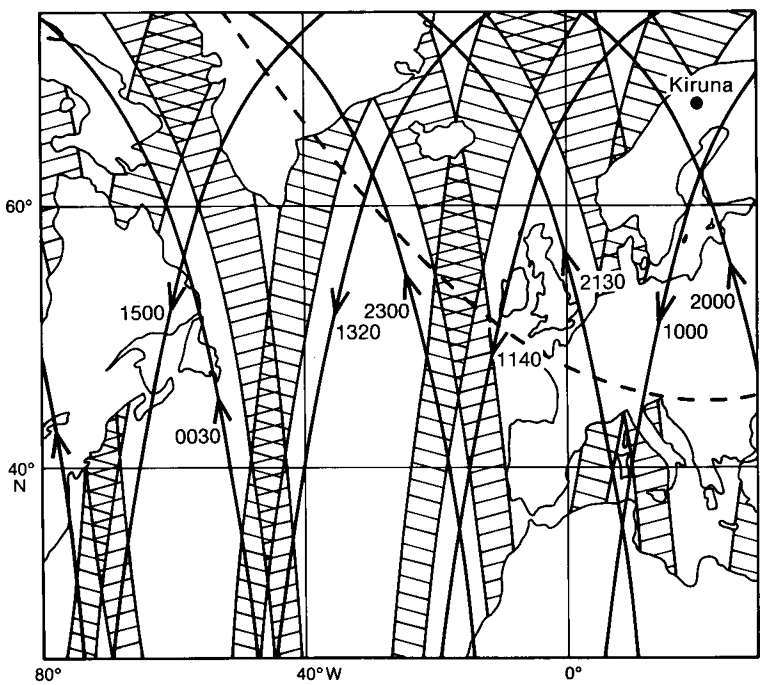

Figure 3. ERS-1 sub-satellite tracks (arrowed lines, with approximate times) and wind scatterometer coverage (hatched areas) of the North Atlantic region over 1 day. The large gaps are partially filled on subsequent days; nominally this occurs on a 3-day cycle. The dashed line shows the limit of reception for the Kiruna ground station.

Ideally, we would like some indication of the 'true' direction, such as from ships or buoys, or, even better, because of greater spatial coverage, the wind analysis from a numerical model. Unfortunately, while these may be available after a few hours, at the time the satellite winds are required these independent data will not be ready. The best that we could have is a forecast field of say 6 or 12 hours ahead for the verification time of the satellite data. Winds from a forecast grid can be interpolated to each scatterometer data point and the solution with the nearest direction chosen. While this can result in high skill scores (over 95%), there is a major drawback; if a forecast is wrong in the positioning of a circulatory feature by 100–200 km (a common occurrence), then there is an area of winds between the forecast centre and true centre which is 180° in error. Here, we will select exactly the wrong scatterometer solution. Worse, if fed back into the next analysis, these winds could reinforce the incorrect forecast which is also being used as the assimilation background. Whilst there are probably enough conventional surface data to prevent this in the North Atlantic region, there could be a problem in the southern oceans and the Pacific where other data are sparse.

5. Current involvement and future use of ERS-1 data

The problems of extracting accurate wind data within the time constraints of an operational system, and the problems of ambiguity removal are the continuing subject of study within the Satellite Meteorology Branch of the Meteorological Office, on behalf of ESA. As part of the ERS-1 project, the United Kingdom is setting up a dedicated off-line ERS-1 data centre, to be constructed at the Royal Aircraft Establishment, Farnborough. This will act as an archive and product-generation facility both for ESA and for the direct needs of the United Kingdom. Here too, the Meteorological Office is involved in developing the algorithms to produce the best possible satellite winds for users not requiring them in real time.

In order to utilize fully the data from ERS-1, ESA has issued an Announcement of Opportunity (AO) for scientists to use the data in operational meteorology, oceanography and for fundamental research. The Meteorological Office has responded positively to this AO, and expects to be a lead agency in assisting ESA to calibrate initially the wind and wave data after launch, and then validate and assimilate the information into its atmospheric and wave forecast models.

The calibration period is expected to be the first 3 months after launch, and will check the correct operation of the hardware and the wind retrieval algorithms. After this period, the model function (the coefficients in equation (1)) may need to be updated. Thereafter, validation — the process of keeping an objective eye on the quality of the satellite-derived data — will be continuous, using a three-way comparison of conventional ship and buoy data, the model analyses (which can be used in data-sparse areas) and the ERS-1 winds and waves. Simulated assimilation using SEASAT winds, currently being undertaken in the Dynamical Climatology Branch of the Meteorological Office, suggests that such satellite data could have most impact when fully utilized as a global data set.

6. Conclusion

The wind and wave data from ERS-1 will not be unique; by the mid 1990s there are expected to be two or three other satellites, operated by the USA, Canada and Japan which could have similar instruments. Indeed, the decision is to be taken soon on whether a follow-on satellite, ERS-2, can be funded within Europe. Further into the decade, the Polar Platform, part of the US Space Station project, is likely to carry similar microwave instruments as part of its fully operational payload. Although ERS-1 is strictly non-operational, ESA intends that it can demonstrate this capability during its 2- to 3-year life. The Meteorological Office, together with other agencies, has taken that challenge and intends to investigate the utility (or otherwise) of this new source of data, exploiting it fully in numerical models and for storm warning and ship routeing services.

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551.524.36:551.588(41-4)

The estimation of extreme minimum temperature in the United Kingdom

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Summary

Current Meteorological Office procedures for the estimation of the lowest temperature to be expected once in 50 years are based on a map produced by Hopkins and Whyte (*Meteorol Mag*, **104**, 1975) in which topographic effects have been removed. More recently Tabony (*J Climatol*, **5**, 1985) quantified some of the effects of topography on minimum temperature, and this paper develops a pragmatic scheme which enables some of his findings to be implemented.

1. Introduction

Hopkins and Whyte (1975) devised a procedure for estimating the maximum and minimum temperatures associated with given return periods by subjecting records from over 200 stations to extreme-value analysis. They produced sea-level maps of the maximum and minimum temperatures to be expected once in 50 years and to which altitudinal lapse rates of 10 °C km^{-1} (for maxima) and 5 °C km^{-1} (for minima) should be applied. The map values are clearly intended to apply to 'standard sites' to which topographical corrections should be freely applied; the authors quote an example of a frost hollow whose recorded extremes are 8 °C below the map value for minima. More recently Tabony (1985) attempted to quantify the effect of topography, and the aim of the present paper is to incorporate his results into an improved scheme for the estimation of minimum temperature. Although the 1985 investigation was limited to inland and rural locations and offered no new information on the effects of water or buildings, the implementation of the findings relating to the lie of the land is still very useful. Many of the requests for estimates of minimum temperature, including, for instance, those concerned with the erection of masts on hills, relate to locations where the effect of topography is dominant.

2. Departures from the Hopkins and Whyte values

The lowest temperature recorded in the United Kingdom, namely −27 °C at Braemar (for location of places in Scotland mentioned in text see Fig. 1), has been observed twice in the last 100 years or so and hence may be accorded a return period not very different from 1 in 50 years. The Hopkins and Whyte map is reproduced in Fig. 2 and it can be seen that the observed temperature is 5 °C below the map value, but as the station is not in the valley bottom it is quite likely that temperatures 8 °C below the map values could be observed in that area. In England the −25 °C reported from Shawbury is 7 °C below the map value and so, with the addition of the Hopkins and Whyte example, there is firm evidence that in the coldest frost hollows, temperatures 8 °C below the map values may be reached.

The lowest temperatures recorded on summits or high slopes, however, are well above the Hopkins and Whyte values, especially when the 5 °C km^{-1} altitude correction is applied. The examples from Scotland shown in Table I illustrate the point. The observed minima have been used to obtain the once in 50-year values by reference to the graph published by Hopkins and Whyte which indicates, for example, that the '20-year' and '50-year' minimum temperatures differ by only 2 °C. It can be seen that the Hopkins and Whyte estimates of minimum temperature on summits are up to 10 °C too low, although the maximum difference is reduced to 7 °C if the altitudinal correction is not applied.

Large lakes which do not freeze over also have a very large warming effect, as witnessed by the lowest recorded temperatures of −12 °C at Sloy on Loch Lomond (1965–78) and −13 °C at Ardtalnaig on Loch Tay (1958 onward). The largest effects are local, as demonstrated by the −17 °C observed at Fort Augustus (1914 onward) and −20 °C at Dall (1961 onward) where the stations are located further from the water's edge (of Lochs Ness and Rannoch respectively). The minima reported from these stations are close to the map values, but it has to be remembered that without the lakes temperatures in these valleys

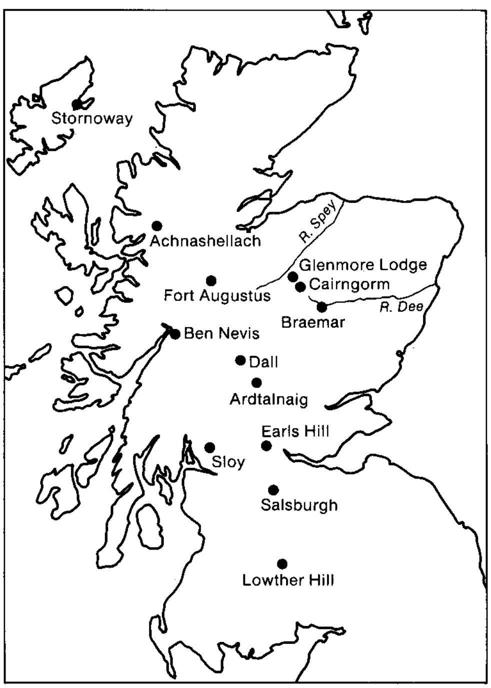

Figure 1. Map showing Scottish stations referred to in the text.

would be several degrees below map values. Valleys which drain directly to the coast are also much milder than those which do not, as exemplified by the modest minimum of −15 °C recorded at Achnashellach (1922–83).

Note that the likelihood of a lake remaining unfrozen in periods of severe weather is related to its depth rather than its area. The cooling of a lake will be related to the ratio of its surface area to its volume, and this will be least for deep, narrow lakes. Thus, large, shallow lakes such as Loch Leven (Kinross) and the southern end of Loch Lomond freeze over, while deep, narrow lakes such as the northern portion of Loch Lomond do not.

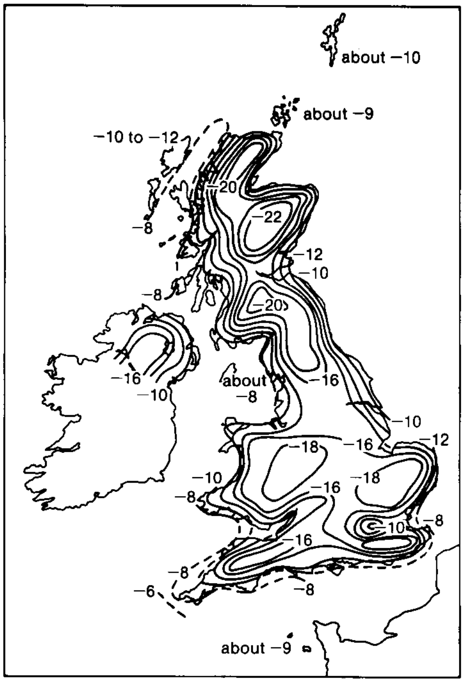

Figure 2. Hopkins and Whyte's map of the annual minimum temperature (°C) to be expected once in 50 years at mean sea level.

Table I. *Comparison of minimum temperatures observed on hills with those estimated by Hopkins and Whyte*

Station	Altitude	Period of record	Station analysis		Hopkins and Whyte	
			Observed minima	1-in-50-year minima	Map value	Altitude corrected
	m		°C	°C	°C	°C
Ben Nevis	1343	1884–1903	−17	−18	−14	−21
Cairngorm Chairlift	1090	1963–1971	−16	−18	−22	−27
Cairngorm Coire Cas	762	1964–1973	−15	−17	−22	−26
Cairngorm Car Park	663	1980 onward	−12	−15	−22	−25
Lowther Hill, Lanarkshire	754	1960–1968	−12	−14	−20	−24
Earls Hill, Stirlingshire	335	1962–1980	−12	−13	−17	−19
Salsburgh (summit of M8 motorway)	275	1965–1974	−11	−13	−17	−18

In the west of Scotland many of the valleys contain lochs, or drain direct to the sea, and this is reflected in the siting of the majority of observing stations. In the east, however, large lochs are rare and hence observations from them are in the minority. Thus we find that the Hopkins and Whyte map has been drawn to fit the −15 °C at Achnashellach and the −12 °C at Sloy whereas the −12 °C at Ardtalnaig is no less than 9 °C above the map value. It is also worthy of note that the coastal gradients drawn on the Hopkins and Whyte map cannot be expected to apply at high altitudes. The lowest temperature recorded at 900 mb by radiosondes released from Stornoway during the period 1961–70 was −13 °C and this is consistent with the −17 °C and −16 °C observed at Ben Nevis and Cairngorm Chairlift respectively.

3. Quantification of topographic effects

Tabony (1985) found it convenient to distinguish between local and large-scale shelter in his description of the effects of topography on minimum temperature. Local shelter can be interpreted as the height above the valley, and large-scale shelter as the depth of the valley. Although attempts to quantify the effects of large-scale shelter were largely unsuccessful it was found that local shelter could be well represented by the maximum drop in height, h, within a radius of 3 km of the station. This gave an objective representation of the potential for drainage of cold air away from the observation site. A relationship of the form

$$\Delta T = A \exp(-0.0055h) \qquad \ldots \ldots \ldots \ldots \ldots \ldots \quad (1)$$

was found, where ΔT is the temperature difference between the free atmosphere and the site (in the absence of large-scale shelter) and A is the temperature difference between the free atmosphere and level ground (i.e. $h = 0$). The temperatures ΔT and A are expressed in °C and h in metres.

The exponent describes the way in which the temperature difference increases as h decreases. Thus, three quarters of the difference between the free atmosphere and level ground has occurred when $h = 46$ m. This was the average value of h for the stations used by Tabony and hence may be taken as the value to be attached to a 'standard site'. Half the difference occurs when $h = 120$ m and less than one fifth when $h = 300$ m.

4. Development of a practical scheme

The *Upper-air summaries* (Meteorological Office 1979–81) reveal only small differences across Great Britain in the lowest temperatures experienced in the free atmosphere. The lowest observed values at 900 mb at Crawley and Stornoway for instance, were −12 and −13 °C respectively, with the corresponding figures for 800 mb being −18 and −20 °C. These values are consistent with a standard lapse rate of 6 or 7 °C km^{-1} which may be extrapolated to give a surface temperature of −7 °C. A couple of degrees can be subtracted to convert from the 10-year extremes of the *Upper-air summaries* to the 50-year return period with which we wish to deal. This gives a value of −9 °C for a free atmosphere or summit value at sea level, if this were possible. This is also consistent with the Hopkins and Whyte figures for the west coast. An altitudinal correction of 7 °C km^{-1} can now be applied to give the summit value at any altitude; this yields −18 °C for Ben Nevis which seems reasonable. Therefore the free-atmosphere temperature, T_f, (°C) is given by

$$T_f = -9.0 - 0.007H \qquad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \quad (2)$$

where H is the altitude in metres.

The Hopkins and Whyte map shows that there is not much geographical variation across the United Kingdom in the 50-year minimum screen temperature at topographically similar sites, and a value of −18 °C may be assigned to level ground. This suggests that the temperature difference between the free atmosphere and level ground is 9 °C, and that this is the value of A in equation (1). Tabony found that, averaged over the whole of Great Britain, the value of A associated with a 30-year extreme in a winter month was 6 °C. There are no problems in raising this to 9 °C to cope with 50-year events over the whole year and in changing from a root-mean-square regression which estimates a mean effect to one which errs on the side of caution (i.e. produces estimates which are too low rather than too high).

So far the effect of large-scale shelter has not been taken into account. The lowest temperatures of −27 °C recorded in the Dee and Spey valleys suggest that 18 °C needs to be subtracted from the free-atmosphere value, i.e. double the level-ground correction. Thus the simplest way of including the effect of large-scale shelter is to multiply A in equation (1) by a factor L which ranges from 1 to 2. The value of L will have to be assessed subjectively but, being a function of large-scale rather than small-scale topography, it is capable of being mapped.

Finally, it is necessary to include the effects of lakes and towns. If these are of any size at all they will counteract any influence of large-scale shelter and also diminish the 'level-ground' effect. Values in the centre of London, for instance, are similar to those experienced on the coast. Urban effects are greatest in districts where the area of sky visible is severely curtailed, i.e. where streets are narrowest and buildings tallest. All these factors can be taken into account by multiplying A in equation (1) by a water/building factor, B, which ranges from 0 to 1.

The procedure outlined above leads to the minimum temperature, T_m, (°C) being given by

$$T_m = T_f - LB\Delta T. \qquad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \quad (3)$$

5. Conclusions

If the site is near the coast the Hopkins and Whyte map is adequate. Otherwise the following procedure may be used:

(a) Obtain the site altitude, H, and the maximum drop in height within 3 km, h, both in metres.

(b) Estimate the free-atmosphere temperature, T_f (°C), from equation (2).

(c) Calculate the effect of local shelter from equation (1) with $A = 9.0$ or interpolate from the following table.

h (metres)	0	50	100	200	300
Local-shelter effect (°C)	9.0	6.8	5.2	3.0	1.7

(d) Subjectively estimate the effect of large-scale shelter as lying between 1 and 2, where a value of 1 is appropriate to level ground and 2 to the maximum known effect in Britain. Multiply the local and large-scale shelter effects to get an overall topographic factor ΔT_t. Reduce the topographic factor according to the presence of lakes and towns (deep lakes and towns can eliminate the topographic factor almost entirely).

(e) Estimate the minimum temperature as the free-atmosphere value obtained in (b) minus the topographic factor arrived at in (d) (see equation (3)).

The steps described above become clearer in the examples presented in Table II.

The fact that the minimum temperature observed at Glenmore Lodge in a (broken) 30-year record was only −20.4 °C serves to illustrate the problems in assigning a value to the large-scale shelter factor. Until this quantity is mapped, the recorded extremes at nearby climatological stations, in combination with a knowledge of the data of those extremes and the station values of h, should enable a suitable value to be chosen. In effect, this means that one will be using the drop in height within 3 km to interpolate between the 50-year minima estimated for individual stations in the local area. Many of the recent enquiries for estimates of extreme minima, however, relate to the erection of masts on hills where temperatures are less sensitive to the effects of large-scale shelter, and where the above procedures should enable reliable estimates to be provided.

Table II. *Examples of the procedure used to estimate the minimum temperature for inland sites (for explanation of symbols see text)*

	Glenmore Lodge	Cairngorm Car Park	Cairngorm Summit
Altitude (H)	341 m	663 m	1245 m
Drop in height within 3 km (h)	22 m	325 m	695 m
Free-atmosphere temperature (T_f) $T_f = -9.0 - 0.007H$	−11.4 °C	−13.6 °C	−17.7 °C
Local-shelter effect (ΔT) $\Delta T = 9.0 \exp(-0.0055h)$	8.0 °C	1.5 °C	0.2 °C
Large-scale shelter factor (L)	2.0	1.0	1.0
Water/building factor (B)	1.0	1.0	1.0
Final topographic effect (ΔT_t) $\Delta T_t = LB\Delta T$	16.0 °C	1.5 °C	0.2 °C
Minimum temperature (T_m) $T_m = T_f - \Delta T_t$	−27.4 °C	−15.1 °C	−17.9 °C

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Reviews

Climate, weather and Irish agriculture, edited by T. Keane. 155 mm × 245 mm, pp. xvii + 187, *illus.* Dublin, Mount Salus Press Ltd, 1986. Price Ir. £9.95 (paperback), Ir. £14.95 (hardback).

Weather affects many agricultural operations and it follows that the proper use of weather information should make it possible to use resources more effectively. This book is intended to encourage a greater awareness of the role of weather and climate in agriculture; it is directed at all involved in agriculture but particularly to students at universities and agricultural colleges.

This is not a textbook about meteorology or instruments or the Irish climate, though it necessarily includes chapters on these subjects. It is a book about the relationship between weather and agriculture. The subjects discussed in the book include: weather and crop production; amelioration of climate for horticultural purposes; animals and the environment; pests and diseases of crops and animals; weather and forestry. Separate chapters are devoted to the moisture balance, to the energy balance and to climate and soil management, and there is a review of the potential of the Irish climate for agriculture. Each chapter contains the necessary biological, physiological and scientific information for an understanding of the general principles involved in the processes under discussion, whether it be the use of mulches to conserve water or the effect of weather on lactation in cows.

The book has been written by specialists 'chosen for their knowledge and experience in the interface area between meteorology and agriculture'. The contributors are from the Irish Meteorological Service, from An Foras Talúntais (the Agricultural Institute), from the University Colleges of Galway and Dublin and from the Veterinary Research Laboratory. Each chapter presents a comprehensive survey of all aspects of the topic under consideration, with references to the operational and research work of other countries, including the United Kingdom. Each chapter has a bibliography of standard reference works and recent research papers and, in total, over 500 books and publications, mostly non-meteorological, are listed.

Several references are made in the book to the work of, and papers produced by, the agriculture section of the Meteorological Office. The reference to the joint Meteorological Office–ADAS (Agricultural Development and Advisory Service) plant disease warning scheme is slightly outdated in that net blotch warnings are no longer issued; the warning criteria were developed using 'in-crop' weather data and gave poor results when used with synoptic data. The Office's crop disease environment monitor (CDEM) mentioned as 'currently being evaluated in Britain' was not a conspicuous success; it suffered major problems with the internal battery pack and was unreliable in operation. The system nevertheless demonstrated the value of in-crop weather data and the need is now being met by commercial firms with more efficient data-logging equipment.

Criticisms can be made of this publication, mostly of a minor nature. Some of the chapters attempt to include too much detail, some of which, though relevant, is neither fully researched nor particularly helpful! For instance, in chapter 10 we learn that '...flights of winged ants tend to occur after rain.' That is the entire scholarship on the subject of winged ants. We learn also that eggs and larvae of wire-worm and leather-jackets are prone to dessication, though the authors admit that dessication is unlikely to be a problem in Ireland. Many other bugs receive an honourable mention and not much else.

The attempts to compare the climate of Ireland with the climate of north-west Europe and with New Zealand in terms of the agricultural potential are unsatisfactory. There was never any possibility that this could be done adequately in two and a half pages and the book would not have suffered in any way for the omission of this small section.

The book is well illustrated with line drawings relevant to the text. Some photographs have been included which show meteorological sites or equipment; they are, however, too small and too general to add anything of value to the book. Photographs of the pests and diseases referred to would have added interest for those on the fringes of agriculture.

The index, which runs to over 3000 items, is less than helpful unless the reader knows precisely what to look for. For example, while there is an entry for nematode worms, the four and a half pages on plant nematodes in chapter 10 are not found under that general heading but under their individual titles, e.g. 'potato cyst nematode'. Similarly, potato blight is known as 'late blight' in Ireland and appears in the index under this title, rather than under 'potato blight' or plain 'blight'; the biological index is useful for those knowing the Latin names of biological organisms.

The suggestion that the book will help producers at all levels to maximize their returns is not easy to justify. No attempt has been made to quantify the cost–benefit of meteorology applied to agriculture and little practical guidance is given on how best to apply the information for planning and scheduling operations. Some specific facts and figures on the costs of the damage caused by pests, diseases and poor weather, together with some clear indication of where and how savings are to be made by the wise use of available weather information would have been helpful.

It is, however, difficult to doubt the other claim — that this book is a valuable addition to Irish agricultural literature. It will also find a place in many UK and overseas libraries. This book is to be commended to all meteorologists involved with agriculture, but especially the newcomer to the field of agrometeorology who will find in it a wealth of useful information and a comprehensive overview of agriculture with weather in its proper perspective.

P. Harker

The uncertainty business, by W.J. Maunder. 161 mm × 242 mm, pp. xxviii + 420, *illus.* London, Methuen and Co. Ltd, 1986. Price £45.00.

The uncertainty business appears at first glance a somewhat daunting book; the text is sandwiched between nearly 30 pages of lists, prefaces and acknowledgements, and 90 pages of indices and bibliographies. At second glance too, as further inspection reveals pages full of the abbreviated names of hundreds of committees and organizations, and sentences so long and tortuous that I had to read them several times over to understand them. Alas, the book doesn't invite further reading!

In the initial section, considering the atmosphere as an elite resource, Maunder asserts that the techniques of assessing the impact of the atmosphere on human activities have been badly neglected, and that such impact has significant economic, social and political consequences. He argues that meteorologists should now take their place alongside other disciplines — economics in particular — to guide important decision- and policy-making in all walks of human affairs.

There follows a comprehensive review of the various developments in monitoring and understanding the effect of the atmosphere on the management of resources. The author argues that the real benefits of weather information will come not only when the right information is produced, but also when that information is used efficiently in decision-making processes at all levels. Indeed, until we know the true impact of tomorrow's weather upon human activities, we are not in a good position to assess the importance and value of forecasting it. This is the important central thrust of Maunder's argument.

He goes on to review the major recent studies assessing the effects of weather and climate. The choice of examples, however, taken mostly from Maunder's own researches in New Zealand, gives a rather insular view. That said, there follows an impressive array of tables and graphs demonstrating the benefits of these methods. Here, Maunder goes for quantity in his arguments, heaping example upon example, giving the reader little opportunity to stand back and admire the view; around page 180, statistical indigestion begins to set in.

This is particularly unfortunate, as Maunder has now come to arguably the most significant part of the book — a series of case-studies, assessing the economic impact of weather and climate upon a range of human activities. Not only are the traditional ones here — agriculture, transport, building construction and energy — but also the new market places which are as yet largely undeveloped meteorologically — road construction, manufacturing and retailing. Many good examples and methods of analysis are given; possibly because these are mostly aimed at identifying the relationship between production and climate variables, rather than simply defining climatological indices, this section is more satisfying and convincing. Maunder demonstrates a variety of techniques of analysis, each to some extent determined by the industry concerned. These studies point to the very considerable benefits which would accrue if meteorological information was applied more efficiently and widely. In essence, this section is a primer for organizations which wish to capitalize on these benefits.

Maunder briefly considers the use of statistical models to predict the likely impact of climate. He proposes developments of this work to include the weather-sensitivity analysis of national economic indicators and their subsequent 'weather adjustment'. Such developments, he says, although possibly controversial, would give economists and politicians a much clearer view of underlying economic trends.

Finally, Maunder identifies the major challenges and opportunities in the management of weather-sensitive resources — the rise of carbon dioxide levels, acid rain, and the nuclear winter. These issues are, perhaps wisely, merely indicated; further discussion is left to other forums.

Taken as a whole, this book raises important issues with which many meteorologists will be familiar, but which need to be debated much more fully outside the meteorological community. The application of the study of weather and climate to all walks of life through assessment of its political, social and economic impact, has a potentially vital role. The book's very comprehensiveness lends it undoubted authority, but this same quality, coupled with its failure to highlight clearly the central themes and conclusions, detracts from its value as a persuasive document. The flyleaf describes it as a manifesto; if so, it should have been far shorter, and written with a far wider audience in mind.

F.R. Hayes

The Irish meteorological service: the first fifty years, 1936–86, edited by L. Shields. 208 mm × 296 mm, pp. viii + 107, *illus.* Dublin, The Stationery Office, 1987. Price Ir. £6.00.

'Unto thine own self be true.' Polonius was not always an old fool and, perhaps, here he was not just being moral but also nurturing Laertes's self-confidence. Being true to itself, self-confidence and active up-to-date maturity are three main characteristics of the Irish Meteorological Service which permeate and shine out from this happy book.

Its title summarizes its scope and objective: to review the work and personalities of the Irish Meteorological Service from 1936 to 1986. It does that, and a great deal more. The front cover sets the scene with the first of a number of excellent and sometimes dramatic colour photographs: does any other Meteorological Service have its home in a frustum of a pyramid? A ziggurat? The back cover has a METEOSAT picture of the globe, which would be commonplace if it were not one for St Patrick's Day. This attention to detail must be one factor in making the Irish Meteorological Service worthy of its high international reputation.

Its people must be another. Its first Director was a former Secretary and Treasurer of the Meteorological Office branch of the Institution of Professional Civil Servants. Its second was a Basque who had fought against Franco in the Spanish Civil War. Subsequent staff have all played their individual parts and many of them are recorded here or have contributed an essay.

The humanity of the people who have served meteorology in Ireland, and continue to do so, must be another secret of the Service's ability to serve its many clients in a wide range of weather sensitive activities: its Minister congratulates them 'with a heart and a half'. It is uplifting to see throughout the book the Service's concern for people, an emphasis on the contribution they make and the way in which their nature determines the success of the organization. 'At times Mr Finnegan worried about the small pension he would qualify for when he retired. He need not have worried. The kindly gentle soul departed this life before he qualified'. Like the Service, the book is also based around people; it is made up of a set of essays by a formidable panel of the Service's experts. They cover every aspect of meteorological life in Ireland, both of yesterday and of today: from the truth behind the great potato blight in 1845 to serving the off-shore oil industry in 1986.

The list of meteorologists from Ireland is brilliant with names such as Robert Boyle, Admiral Beaufort, Robinson (of cup anemometer fame), Stokes (Equation and sunshine recorder), Sabine, Tyndall and R.H. Scott (Director of the Meteorological Office). To quote names of the living would be invidious; but it must be said that the forecasting talent brought together at Foynes to support the beginnings of transatlantic air travel can rarely, if ever, have been matched.

Today the Service comes over as up to date, making good use of the data and products it receives, and contributing to meteorology in other countries, not only through participation in the European Centre for Medium Range Weather Forecasts and the usual exchange of World Weather Watch data but also in many other ways. Two instances are the graphics software it has developed in-house and its training programme, both made available through the Voluntary Co-operation Programme of the World Meteorological Organization. In two articles, the book distinguishes carefully between the accuracy of forecasts and their utility: the Service is not only human, it is professional.

Until 1936 the responsibility for meteorology in Ireland lay with the Meteorological Office, so that we may take pride in having a thriving offspring, which has grown to full maturity and continues to have a full and friendly relationship with us. In Hamlet, Laertes dies. In Ireland he has flourished into his second half-century of confident well-balanced maturity and has published a most readable autobiography.

To summarize the book or the history of the Service here might satisfy your curiosity. That would be a shame. You would do better to beg, borrow or even buy — and then enjoy — a copy for yourself.

S.G. Cornford

Landolt-Börnstein: numerical data and functional relationships in science and technology, V/4a, edited by G. Fischer. 200 mm × 277 mm, pp. xii + 491, *illus.* Berlin, Heidelberg, New York, London, Paris, Tokyo, Springer-Verlag, 1987. Price DM 1220.00.

This volume is part of a new series of Landolt-Börnstein books which give exhaustive compilations of numerical constants, tabulated data and functional relationships (i.e. equations etc.) in the physical sciences. At present, these are broadly classified into six groups (e.g. Group I is Nuclear and Particle Physics, Group II is Atomic and Molecular Physics); Volume 4 of Group V (Geophysics and Space Research) deals with meteorology and is split into three subvolumes — the first of which is reviewed here.

As the editor admits in the preface, the choice of material covering the dynamics and thermodynamics of the atmosphere is 'a little problematic'. Furthermore the Landolt-Börnstein philosophy of presenting just tables and figures is transgressed and substantial space is given to theoretical discussion. One can find, for example, a crude scale-analysis of terms in the equations of motion, a review of the Lorenz concept of available potential energy and even a short history of the European Centre for Medium Range Weather Forecasts (ECMWF).

The first section of the book gives a compressed account of the equations of motion in various co-ordinate representations together with some commonly used reduced forms of these equations such as the barotropic and quasi-geostrophic models. I could not find any reference to the important Boussinesq approximation or the semi-geostrophic equations; neither is the concept of potential vorticity given anything more than a brief mention. Although the presentation is clear, much of the material is dated and readily available elsewhere.

The second section, on atmospheric data, provides a wide range of information including tables for the conversion of degrees Fahrenheit to Celsius(!), geostrophic wind speed from isobar spacing, and virtual temperature increments as a function of pressure and temperature. It also contains a thorough treatment of the physical properties of moist air, with many obscure and complicated formulae. I suspect that much of the information given in this section will not be sought after, e.g. tables for the wavelength of the stationary external Rossby wave as a function of latitude and zonal wind speed, and formulae for the vertical structure of polytropic model atmospheres.

Section three, on the general circulation of the atmosphere, forms over one half of the book and contains a collection of time-mean diagnostics derived from 17 years of National Meteorological Center (NMC) analyses. Time-averaging is taken over the four seasons separately for both hemispheres. For instance, one can find maps of the horizontal distribution of temperature at 850, 700, 500, 300, 200, 100 and 10 mb as well as meridional cross-sections of the spectral amplitude (though not phase) of temperature and height (of pressure surfaces) for wave numbers 1, 2 and 3, and various spectral decompositions of the horizontal transport of heat and momentum. Much of the zonally averaged and spectral data are also tabulated even when, as in Table 25e on page 387 (the inter-annual variance of eddy heat transport in wave number 3), all entries in the table are zero! The data of this section will nicely complement existing general circulation statistics (e.g. Oort 1983)*.

Finally, there is a section dealing with meteorological organizations providing, for example, the history and structure of the World Meteorological Organization and an address for virtually all the national meteorological services world-wide.

There is no doubt that the book contains a lot of useful information, but is it really worth £430? Much of the tabulated data is either available in other textbooks, obtainable on magnetic tape from NMC or ECMWF, or easily computed on a pocket calculator.

G. Shutts

Books received

The listing of books under this heading does not preclude a review in the Meteorological Magazine at a later date.

Acidification of freshwaters, by M. Cresser and A. Edwards (Cambridge University Press, 1987. £19.50, US $34.50) examines the numerous interacting physical, chemical and biological processes which regulate acidity in freshwaters. Concepts from many disciplines are brought together in an attempt to make a coherent picture which is understandable to a wide range of people.

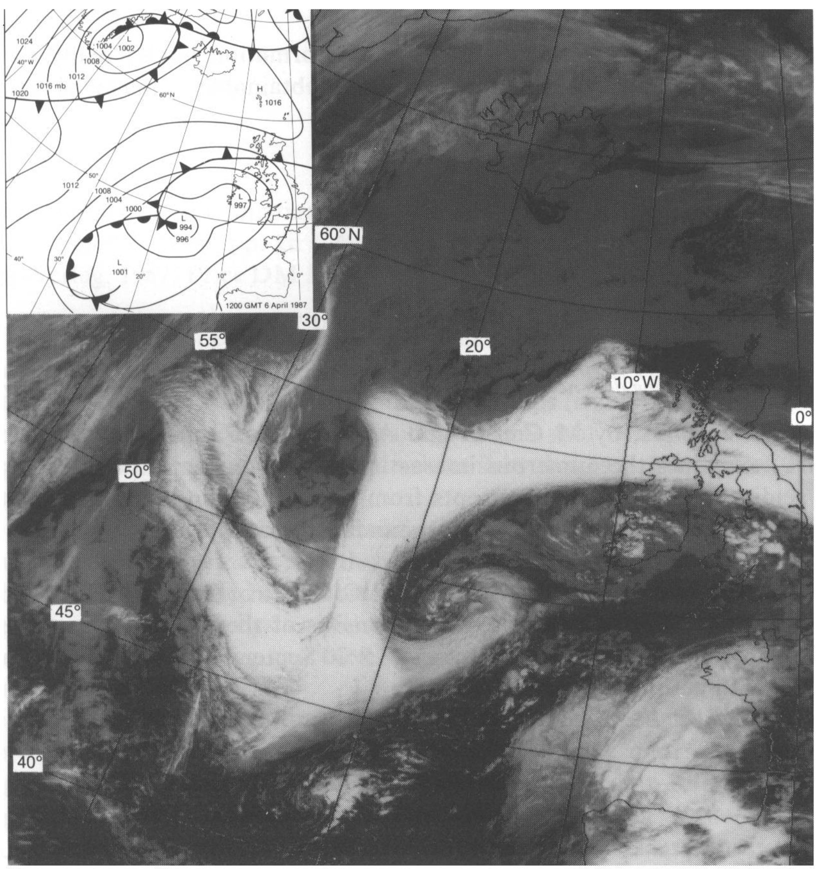
Acidic precipitation, Parts 1 and 2, edited by H.C. Martin (Dordrecht, D. Reidel Publishing Company, 1987. £196.00, US $240.00, Dfl.560.00) consists of the proceedings of the international symposium on the subject, held at Muskoka, Ontario, 15–20 September 1985. There are over 200 papers on related subjects, with author and subject indexes.

* Oort, A.H.; Global atmospheric circulation statistics, 1958–1973. Washington D.C., NOAA, 1983, Professional Paper 14.

Satellite photograph — 6 April 1987 at 1548 GMT

Throughout the period 5–7 April 1987, a complex depression was situated to the south-west of Ireland and a cold front moved erratically and very slowly north over the British Isles. This front was characterized in the infra-red Meteosat pictures by a distinctive band of cloud with apparently uniform brightness and definite edges. The associated surface weather consisted of rain and drizzle, slowly dying out, with hill and coastal fog, mainly in the east. The accompanying picture, which was taken in the infra-red, shows more detail within the cloud band. From a combination of Meteosat images, the wave-like structures on the northern (anticyclonic) side of the front could be seen to move westwards, each developing differently. The cloud streaming north-west of 46° N, 23° W shows the remains of the crest of the first wave, which had crossed Scotland about 30 hours earlier. The crest of the second wave, at 54° N, 22° W, is beginning to curve forwards while the third wave, at 57° N, 12° W, is 'rolling' backwards.

The areas of showers over the west of England, Wales and southern Ireland are associated with cold pools in a band of cold air which extended from Poland to the Azores. The area of cloud at 42° N, 21° W appears like a PVA (Positive Vorticity Advection) area, which soon became absorbed into the re-invigorating main vortex. The area of cloud over the Bay of Biscay swept northwards and brought extensive moderate to heavy rain over England and Wales from the early hours of April 7.



Photograph by courtesy of University of Dundee

Meteorological Magazine

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