Long-Range Forecasting and Climate Research

The climate of the world

I — Introduction and description of world climate

by

C.K. Folland

LRFC 1

March 1986

MO13

LONG RANGE FORECASTING AND CLIMATE RESEARCH MEMORANDUM NO. 1
(LRFC 1)

THE CLIMATE OF THE WORLD

I-INTRODUCTION AND DESCRIPTION OF WORLD CLIMATE

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C K Folland

BASED ON TWO ADVANCED LECTURES DELIVERED TO THE SCIENTIFIC OFFICERS' COURSE, METEOROLOGICAL OFFICE COLLEGE, MARCH 1985

Met O 13 (Synoptic Climatology Branch)
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SZ

March 1986

This Memorandum is the first in a series of six on:

THE CLIMATE OF THE WORLD

BY

C K Folland and D E Parker

Based on nine Advanced Lectures delivered by C K Folland to the Scientific Officers' Course 1-7 March 1985, and one Advanced Lecture delivered by D J Carson in March 1982.

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LECTURE 1 INTRODUCTION TO WORLD CLIMATE

by C K Folland

Summary

This introductory lecture discusses the physical approach to climate studies. The WMO World Climate Programme (1980-2000) is introduced; this program already provides a focus for many activities relating to climate research and the use of climate knowledge and data for practical purposes.

1.1 Introduction

There are several complementary approaches to the study of climate. A traditional image of climate is the one often presented at school. In the preface to his book 'Climatology' (9th Edition 1961) the famous Reading University geographer Professor Austin Miller wrote 'The nature of the relief, the vegetation, the habits of the people, their architecture and their occupations will be attuned to the prevailing rainfall and temperature conditions and it is the response rather than the cause which the geographer wishes to know'. This is still the main approach of the many 'human' geographers who wish to use climate data to explain the development of society and industry in the context of the physical environment. It is nevertheless a legitimate use of climate data, paralleled by the wide-spread use of climate data by industry and agriculture for planning. Here understanding how climate operates is much less important than the goal of collecting and validating climate data to provide reliable analyses of that data tuned to user's needs. However a qualitative understanding of climate is still very desirable when deciding on the choice of appropriate statistics and of suitable data; nowadays many of the analyses needed are far removed from the calculation of simple climatic averages. This 'applied' climate activity is important in several branches of the Meteorological Office.

Over the last few decades, research into the physical basis of climate and more recently its variability has become increasingly feasible and productive due to developments in meteorological theory, computers and thus in meteorological models. In addition, our ability to collect and validate reliable worldwide data has considerably increased. Research into the mechanisms and behaviour of climate is essential if useful climate predictions are to become possible, especially the prediction or simulation of the consequences of increased carbon dioxide and the persistent and socially disruptive droughts endemic in parts of the tropics. In the following lectures most emphasis will be placed on the physical approach to climate.

1.2 What is climate?

The Scientific Plan for the World Climate Research Programme (WMO (1984a)) provides several useful definitions:-

- Weather is the manifestation of a state of the atmosphere at a particular instant in time and the evolution of this state through the generation, growth and decay of individual transient (synoptic scale) disturbances.

- Climate is the synthesis of weather over the whole of a period long enough to establish its statistical ensemble properties (mean values, variances, probabilities of extreme events ...) and is largely independent of any (particular) instantaneous atmospheric state. A climatic state is described by these statistical properties.

- Climate change (or climate signal) is measured by the difference between mean values of a climatic parameter or statistic, i.e. the difference between climatic states.

- Climate variability is measured by the deviations of monthly, seasonal and annual values from the mean value of the climatic state established over an ensemble of such periods. The differences from the mean associated with an individual event are called anomalies.

For the purposes of these lectures a given 'climate' would normally be measured over a period of decades (30 years is a 'standard' period). Climate variability is in principle a manifestation of the variations of time-averaged weather that contribute to a given climate eg the fact that last January was markedly different to the previous January would usually reflect 'variability' but if the average temperature for the last 30 Januarys was markedly different from the average temperture of the preceeding 30 Januarys, we could legitimately talk about climate change. Climate 'variability' of course includes a stochastic component due to transient variations associated with weather.

1.3 The World Climate Programme

In 1974 the Executive Committee of WMO decided that it was important for WMO to take a lead in promoting studies of climate changes and their impact on the natural environment and on food production. So, jointly with ICSU, the International Council of Scientific Unions, they arranged a high level scientific and technical World Climate Conference in 1979 which represented the climate sensitive branches of national economies including agriculture, energy, water resources, fisheries and health (WMO (1979)). They reviewed knowledge of climatic change and variability and their implications for human activities. There followed an outline plan and basis for the World Climate Programme (WCP) 1980-1983 (WMO 540) (1980).

WMO's role is to support and help co-ordinate national activities, not to run them. They have set up a World Climate Programme Office within the Office of the WMO/ICSU Joint Planning Staff at Geneva and have established a Joint Scientific Committee (JSC) with ICSU to provide, amongst other things, scientific guidance for the international aspects of the World Climate Research Programme which is a sub-unit of WCP. The JSC determines the main research objectives and facilitates the exchange of information between scientists.

The World Climate Programme is divided into 4 main sections:-

a. World Climate Data Programme (WCDP)

It might be thought that the voluminous data being collected to support weather forecasting should also be sufficient for climate studies. Indeed the data collected for weather forecasting including that archived in the form of initialised meteorological fields is increasingly valuable for climate studies. Nevertheless elucidation of many of the processes essential to understanding climate fluctuations or climate change as well as monitoring how the climate is currently behaving require much more atmospheric data from above the oceans and remote land areas than are presently available. Even more importantly, reliable worldwide measurements of data from the ocean surface and from the depths of the ocean are badly needed. The study of climate change demands long, homogeneous, records. If we are to make more progress with observational studies of climate change not only must the future coverage of atmospheric, land surface and ocean data be more complete but we must be able to make use of existing historic data. Much of this is not at present easily accessible nor has it been properly validated. So major goals of WCDP are improvement of future observational coverage of the atmosphere and the boundary surfaces and here satellite data will be vitally important. Also, it is hoped to organise the identification and assembly of lengthy climate data sets covering the last 100 years and as complete a set as possible of atmospheric, ocean and land-surface data for about the last 30 years. Figure 1.1 from (WMO (1983)) gives an indication of the proposed management of this program.

b. World Climate Applications Programme (WCAP)

Specific goals of WCAP are to enlighten people and policy-makers concerning the value of applying climate data to socio-economic problems and to ensure this knowledge is available. Examples include agriculture, water, energy and health. In fact such applications are already quite highly developed in some countries eg in UK. In UK climate data is extensively used to aid agriculture eg to advise on irrigation and the need for preventive measures for disease control, to design oil rigs to guard against extremes of wind and waves and to design and simulate the behaviour of storm water sewers and culverts in a cost-effective way etc. Clearly these activities are appropriate in many parts of the world. Figure 1.2 provides an example of a practical application of climate data to agriculture; Fig 1.2 can be used, with additional information, to decide where in England and Wales it may be cost effective to grow maize to make a form of silage to feed animals in the winter.

c. World Climate Impact Studies Programme (WCIP)

This is a more open ended program that seems to come nearest to the 'geographers' approach to climate. The main objectives are to study the impact of climate variability and climate changes on the main economic activities in society (eg agriculture, transport, land use and water resources) and the way that society reacts to these fluctuations with emphasis on the range of impacts in different human societies throughout the world. These societies might be chosen for their different levels of development and different natural environments.

d. World Climate Research Programme (WCRP)

This is the most important unit of the World Climate Programme from the point of view of these lectures. This programme developed from the Global Atmospheric Research Programme of the 1970's. Its main objectives (see WMO (1984a)) are bold: to determine to what extent climate can be predicted and the extent of man's influence on climate. The primary approach for meeting these objectives is based on the use of physical-mathematical models (eg general circulation models) capable of simulating and eventually (perhaps) predicting climate fluctuations and changes. The programme is divided into three streams:-

(1) Stream 1

This stream is meant to establish the basis for long-range weather forecasting ie the prediction of weather anomalies on time scales of 1-2 months (perhaps a little more). Thus the development of long-range weather forecasting is now elevated in an international context to a primary concern of meteorology and climatology. For those interested in more detail, WMO (1984(b)) gives an up to date picture of developments and further references.

(2) Stream 2

This stream aims at predicting variations of global climate over periods of several years. This is an even bolder aim and a special sub-programme of the World Climate Research Programme called TOGA (Tropical Oceans and Global Atmospheric Research Programme) has been set up which formally commenced on January 1st 1985. This programme recognises the crucial role that anomalies in the circulation and heat content of the tropical oceans have on interannual climate variability. It also recognises the fact that variations in the temperature of the upper layers of the tropical oceans, for example, can to some extent be understood without attempting to model the remainder of the global ocean in full detail. This strategy is a consequence of current ideas about the dynamics of the tropical oceans. Here the Coriolis force is weak giving rise to several dynamical processes which differ from those in mid latitudes and whose purely marine effects tend to be confined to the tropics.

(3) Stream 3

This stream aims at studying the patterns of climate fluctuation and variation over several decades or more and assessing the potential response of climate to man-made or natural influences. For example, these influences might be the increase of carbon dioxide in the atmosphere or slow changes in the large scale global ocean circulation. Thus a key sub-programme will be the World Ocean Circulation Experiment (WOCE) planned for the 1990's.

Figures 1.3 to 1.5 give examples of why these three research activities are challenging and interesting. Figure 1.3 shows that nearly 50% of the detrended variance of temperatures in Central England (when the annual cycle is removed) in a recent 5 year period occurred on time scales of 10 days to 2 months. Can we predict some of this 'Stream 1' variance? There are many commercial and industrial concerns who would find it useful if we could do this. Figure 1.4 shows that fluctuations in many meteorological parameters occur in the Pacific region which are coherent from season to season and year to year and which fluctuate on time scales of a few years? Can we predict these 'stream 2' time-scale fluctuations. These may be linked (lecture 5) to many irregularly recurring tropical anomalies, including drought in east Australia and floods in coastal Equador and north Peru. Figure 1.5 shows that average Northern Hemisphere near surface land temperatures have almost certainly fluctuated appreciably over the last century (stream 3). There have also been roughly parallel quite large variations in atmospheric circulation patterns and in ice cover in some regions. Is this a real effect? If confirmed these fluctuations need to be understood if we are to predict the consequences of say large increase in carbon dioxide over the next century with any accuracy. A stimulating introduction to the three streams of the WCRP is given in chapter 1 of 'The Global Climate' edited by J T Houghton (1984).

1.4 Relationship between climate and general circulation

I shall frequently refer to the Met O 20 advanced lectures on General Circulation. There is no hard and fast rule for the difference between the two sets of lectures; however Tony Slingo's and Andrew Lorenc's lectures emphasise the physical causes and principles underlying atmospheric circulation. My lectures will have a similar background but are more concerned with the observed behaviour of statistical ensembles of weather events (climate) and the reasons for that behaviour. I will also have an eye on the question: how could we predict these variations? So my lectures will emphasise the near surface environment that affects man, animals and plants. Of course the atmosphere is 3-dimensional and discussion of 'higher' climatic matters will be brought in as necessary.

1.5 Methods of studying climate

(a) Empirical methods

Most (but not all) scientific knowledge is based on observations. We usually need to make observations to develop theories or choose between competing alternatives. Much empirical work on climate is basically 'detective' work which attempts to find out how the atmosphere and oceans behave on longer than synoptic time scales as few theories yet exist for such behaviour. Outstanding examples of important climate fluctuations which have been discovered empirically are the remarkable 17 year sequence of droughts in the Sahel region of Africa (1968-date) whose continued persistence since the well known initial drought sequence of 1968-1974 was only verified in 1982 (Lamb (1982)) but whose consequences are now painfully obvious (Figure 1.6). On a shorter time scale there is a tendency for the extratropical north Pacific/north American/subtropical west Atlantic atmosphere to adopt a special 'PNA' pattern of variation on time scales of weeks in the winter half year. This variation has an equivalent-barotropic structure in the vertical and seems to be distinct from synoptic scale variability as it exhibits the characteristics of a stationary wave with a non-zonal orientation (Figure 1.7). This atmospheric 'regime' was discovered about 50 years ago by Walker and Bliss (1932), confirmed quite recently by Wallace and Gutzler (1981) and is now found to exist in several atmospheric general circulation models (AGCMs) eg Qiu and Esbensen (1984). The PNA pattern seems to be one of several internal atmospheric modes that can be self stimulated within the atmosphere or in AGCM's but might also be preferentially stimulated (or suppressed?) by external forcing eg due to sea surface anomalies etc (Fig 1.7). These wave-like patterns are sometimes called teleconnections as the fluctuations of a given variable at a selected location eg 500 mb height, tend to be persistently positively or negatively correlated with those at selected distant locations. Teleconnections have long been used in long range forecasting.

1.6 Synoptic climatology

This is a branch of climatology which up to recently has been the province of the empiricist. There are many definitions of the subject but in essence it tries to provide explanations or predictions of regional or local climates in terms of large scale fluctuations of the atmosphere. Only one book purely devoted to synoptic climatology has been published so far (Barry and Perry (1973)), written before the use of AGCM's started to make an impact on the subject. The 'synoptic method' involves the study of changes in circulation patterns measured at fixed times or averaged over fixed intervals. It can be quite a powerful technique when used in a discriminating way; thus it provided the foundations of practical weather forecasting prior to the extensive use of mathematical forecasting models. One prominent technique of synoptic climatology is to classify time-averaged circulation patterns into 'types' which have markedly different consequences for regional surface weather or reflect (sometimes only hopefully) different physical generating mechanisms. The PNA pattern mentioned above would be a good example of a 'circulation type'. Other examples of circulation types include the 4 basic 'elementary circulation mechanisms' of Dzerdzeevski (with 41 sub-units) (see p164 of Barry and Perry (1973)) that attempt to summarise different patterns of extratropical Northern Hemisphere circulation. These patterns were developed to help long-range forecasting in Russia (Figure 1.8). A local example is the catalogue of the so called 'Lamb' types of daily surface circulation patterns over UK whose aggregate over long periods has often been found useful in climate studies and in long-range forecasting for UK (Lamb (1972)). Figure 1.9 gives examples of Lamb types.

1.7 Climate Modelling

This subject has been extensively discussed in the last few advanced lectures on general circulation. A recent development has been the interactive use of synoptic climatological methods and AGCM's to study short time scale climate fluctuations. The best example so far is the ongoing, extensive, international study of the El Nino/Southern Oscillation (eg Nihoul (Ed.) (1985)) where AGCM and OGCM (ocean GCM) experiments have been firmly guided by a series of climatological investigations (eg Rasmusson and Carpenter (1982)). However some of the results of the GCM experiments already go beyond what would soundly be deduced from available observations (see Lecture 5 in this series). In this way the modelling results stimulate new ways at looking at the observations allowing interactive studies to flourish. This is the approach increasingly being adopted in Met O 13.

1.8 Summary

I have tried to introduce the vast range of climate studies and why they are important. In the next lecture we shall explore in outline present-day climates of the major regions of the world including some features of their interannual variability.

LECTURE 1 REFERENCES

Barry, R.G. and Perry, A.H.	(1973)	Synoptic Climatology. Methuen.
Hough, M.N.	(1978)	Mapping areas of Britain suitable for maize on the basis of Ontario Units. ADAS Q. Rev., 31, Winter 1978, pp.217-227.
Houghton, J.T.(Ed)	(1984)	The Global Climate. Cambridge University Press.
Jones, P.D. and Kelly, P.M.	(1983)	The spatial and temporal characteristics of Northern Hemisphere surface air temperature variations. J. Clim., 3, pp.243-252.
Lamb, H.H.	(1972)	British Isles Weather types and a register of the daily sequence of circulation patterns 1861-1971. Met. Off. Geophys. Mem. Vol. XVI, No. 116. HMSO.
Lamb, P.J.	(1982)	Persistence of Sub-Saharan drought. Nature, 299, pp.46-48.
Miller, A.A.	(1961)	Climatology. Methuen.
Nicholson, S.E.	(1985)	J. Clim. Appl. Met., 24. pp. 1388-1391.
Nihoul, J. (Ed)	(1985)	Coupled Ocean-Atmosphere Models. Proceedings of 16th Liege Conference on ocean Hydrodynamics (1985). Elsevier.
Qiu G-Q and Esbensen, S.K.	(1984)	Teleconnection patterns simulated by the OSU AGCM. Climatic Res. Inst. Report No. 51. Oregon State University, Corvallis, Oregon.
Rasmusson, E.M. and Carpenter,T.H.	(1982)	Variations in Tropical Sea Surface Temperature and Surface Wind Fields associated with the Southern Oscillation/El Nino. Mon. Weather Rev., 110, pp.354-384.
Walker, G.T. and Bliss, E.W.	(1932)	World Weather V. Mem. Roy. Meteor. Soc., 4, pp 53-84.
Wallace, J.M. and Gutzler, D.S.	(1981)	Teleconnections in the geopotential height field during the Northern Hemisphere Winter. Mon. Weath Rev., 109, pp 784-812.

WMO (1979) Proceedings of the World Climate Conference, Geneva, 12-23 February 1979. WMO No. 537.

WMO (1980) Outline plan and basis for the World Climate Programme 1980-83. WMO No. 540.

WMO (1983) Plan for the World Climate Data Programme. WCP-59.

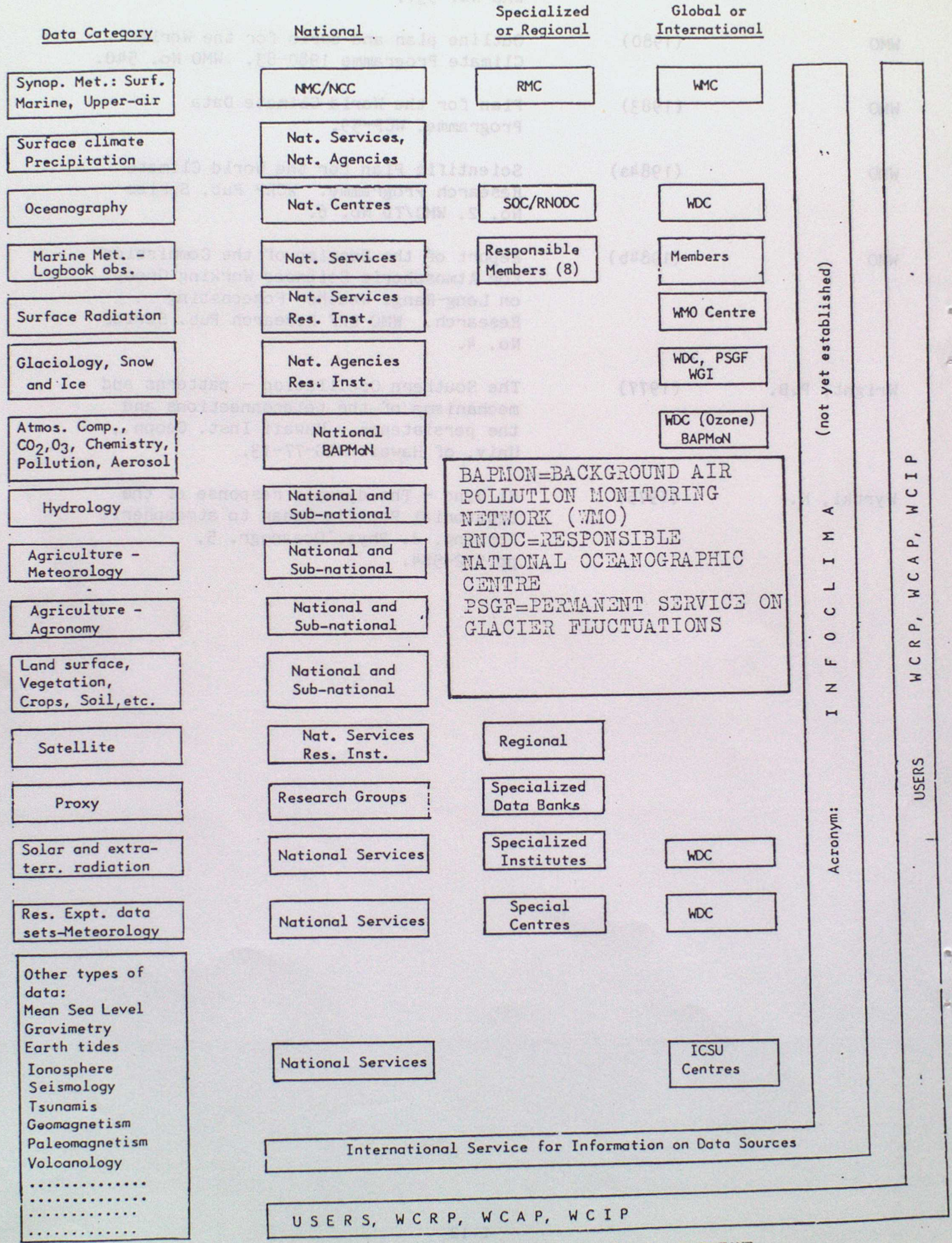
WMO (1984a) Scientific Plan for the World Climate Research Programme. WCRP Pub. Series No. 2. WMO/TD No. 6.

WMO (1984b) Report of the Session of the Commission for Atmospheric Sciences Working Group on Long-Range Weather Forecasting Research. WMO LRF Research Pub. Series No. 4.

Wright, P.B. (1977) The Southern Oscillation - patterns and mechanisms of the teleconnections and the persistence. Hawaii Inst. Geoph., Univ. of Hawaii HIG-77-13.

Wyrtki, K. (1975) El Nino - The dynamic response of the Equatorial Pacific Ocean to atmospheric forcing. J. Phys. Oceanogr. 5, pp.572-584.

Figure 1.1 Data management and Archiving Centres.
Building Blocks of a Co-ordinated Distributed Climate Data Management System
From WCP-59 (WMO(1983))



NMC=NATIONAL METEOROLOGICAL CENTRE WDC=WORLD DATA CENTRE
NCC=NATIONAL CLIMATOLOGICAL CENTRE WGI=WORLD GLACIER INVENTORY
RMC=REGIONAL METEOROLOGICAL CENTRE SOC=SPECIALIZED OCEANOGRAPHIC CENTRE

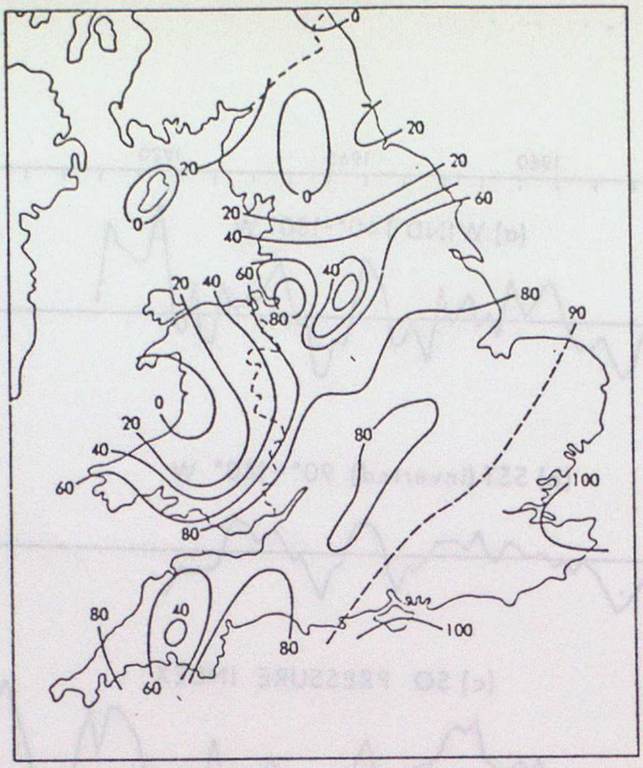


Figure 1.2 Percentage of years suitable for making Tower Silage from maize in England, Wales and the Isle of Man.(From Hough(1978))

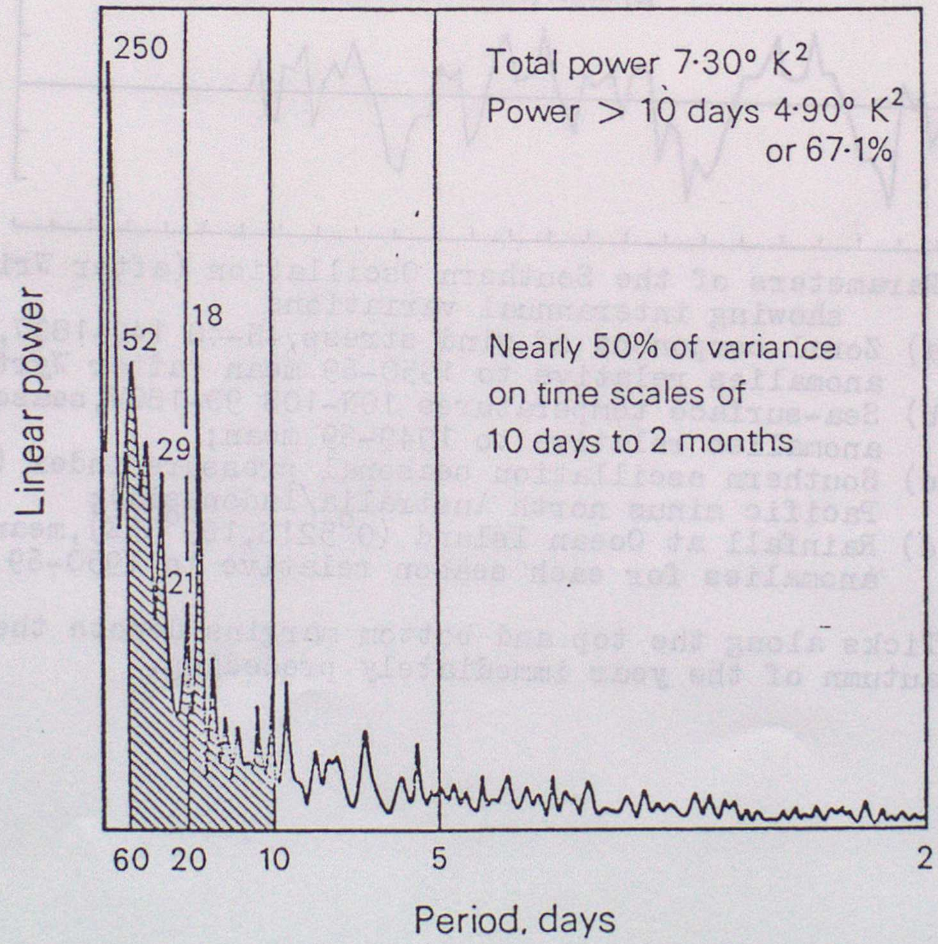


Fig. 1.3 Power spectrum of daily central England temperature (1976-1980) with annual cycle removed

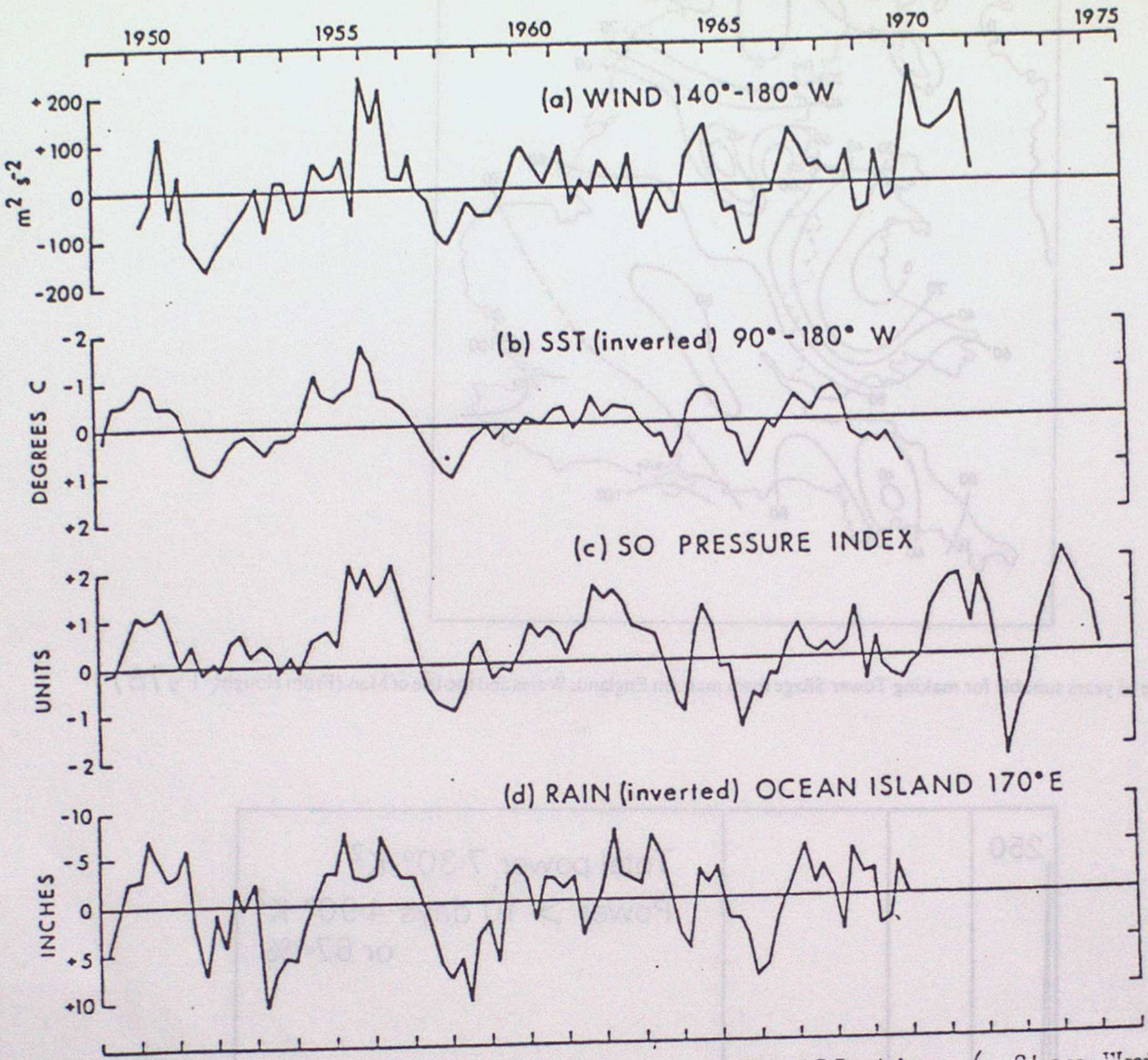
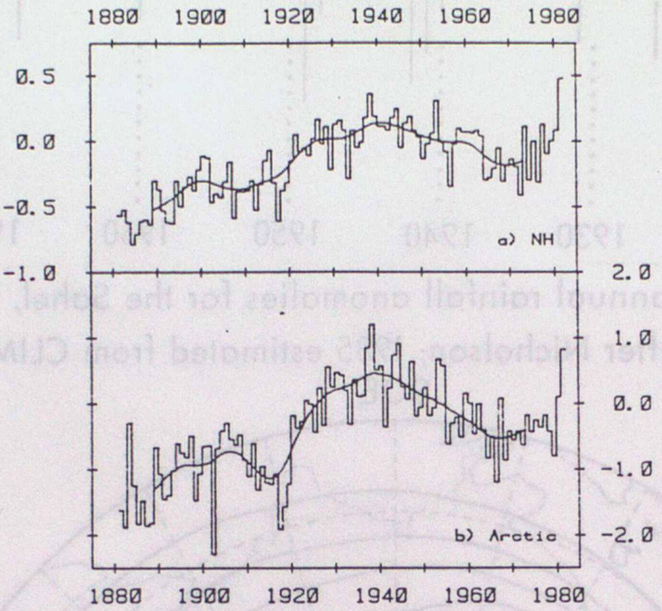


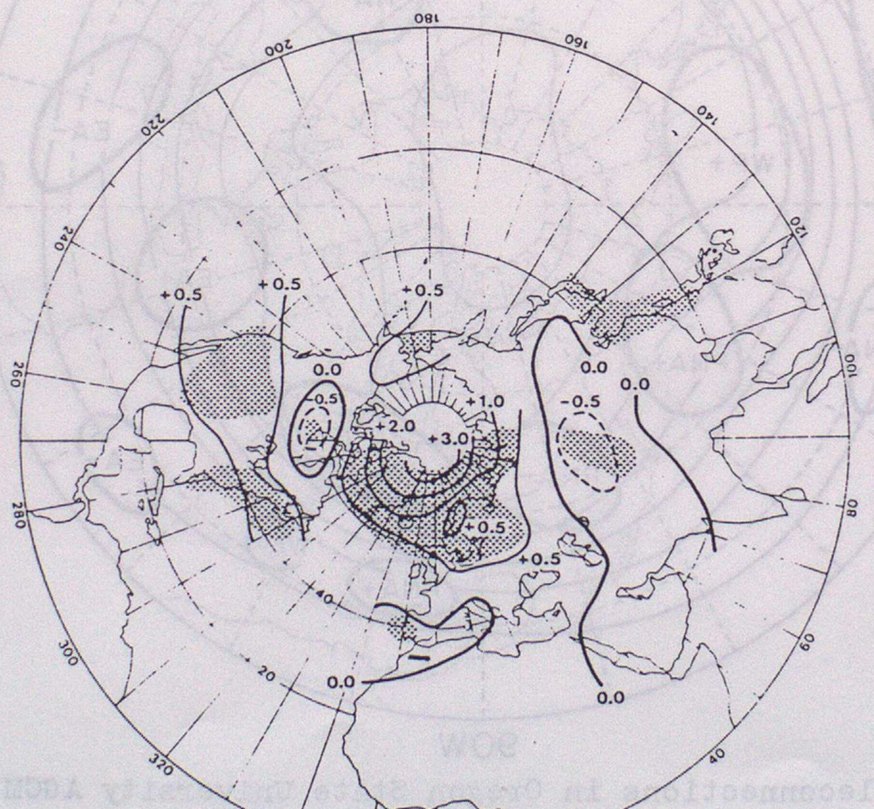
Figure 1.4 Parameters of the Southern Oscillation (after Wright(1977)) showing interannual variations

a) Zonal component of wind stress, 4N-4S 140-180W, seasonal anomalies relative to 1950-69 mean (after Wyrtki(1975));
b) Sea-surface temperatures 10N-10S 90-180W, seasonal anomalies relative to 1949-69 mean;
c) Southern oscillation seasonal pressure index (southeast Pacific minus north Australia/Indonesia);
d) Rainfall at Ocean Island (0°52'S, 169°35E), mean monthly anomalies for each season relative to 1950-69 mean.

Ticks along the top and bottom margins denote the northern autumn of the year immediately preceding.



A Annual means of surface air temperature for the Northern Hemisphere and for the Arctic (expressed as departures in °C from the 1946–1960 reference period). Data for 1981 are included in this figure but have not been analyzed. The smoothed curves show the data smoothed with a binomial filter designed to suppress variations on time scales of less than 20 years



B Linear trend of gridpoint temperatures over the period 1917–1939 (°C/y × 10^{-1}). Shaded areas are significant at the 5 per cent level

Figure 1.5

After Jones and Kelly (1983)

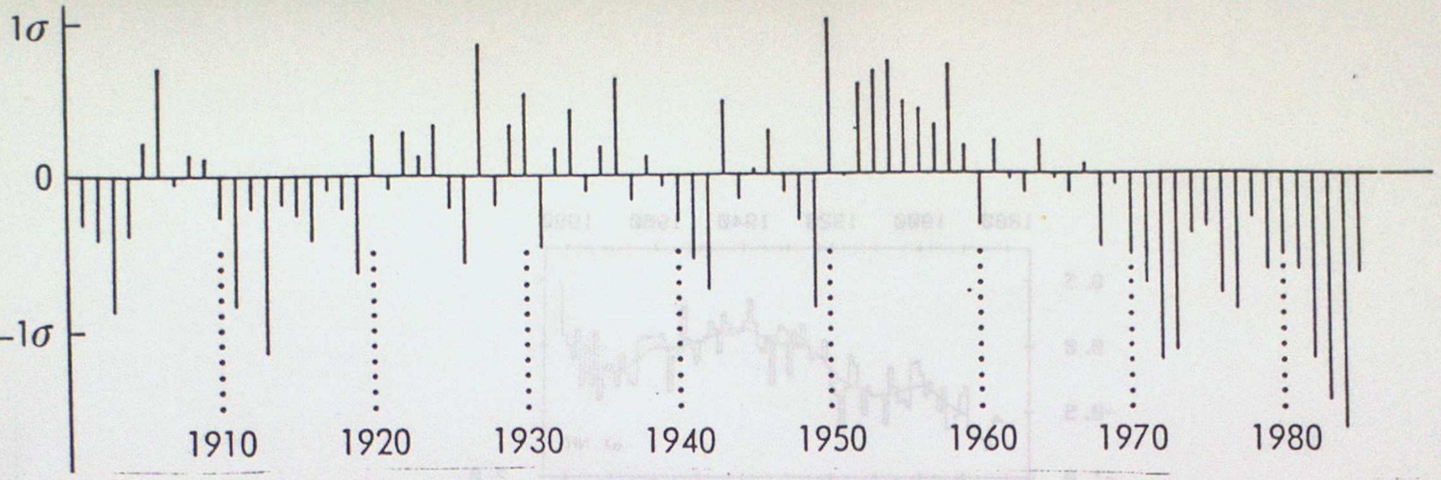


Figure 1.6 Standardised annual rainfall anomalies for the Sahel, 1901—85. Values to 1984 after Nicholson; 1985 estimated from CLIMAT reports.

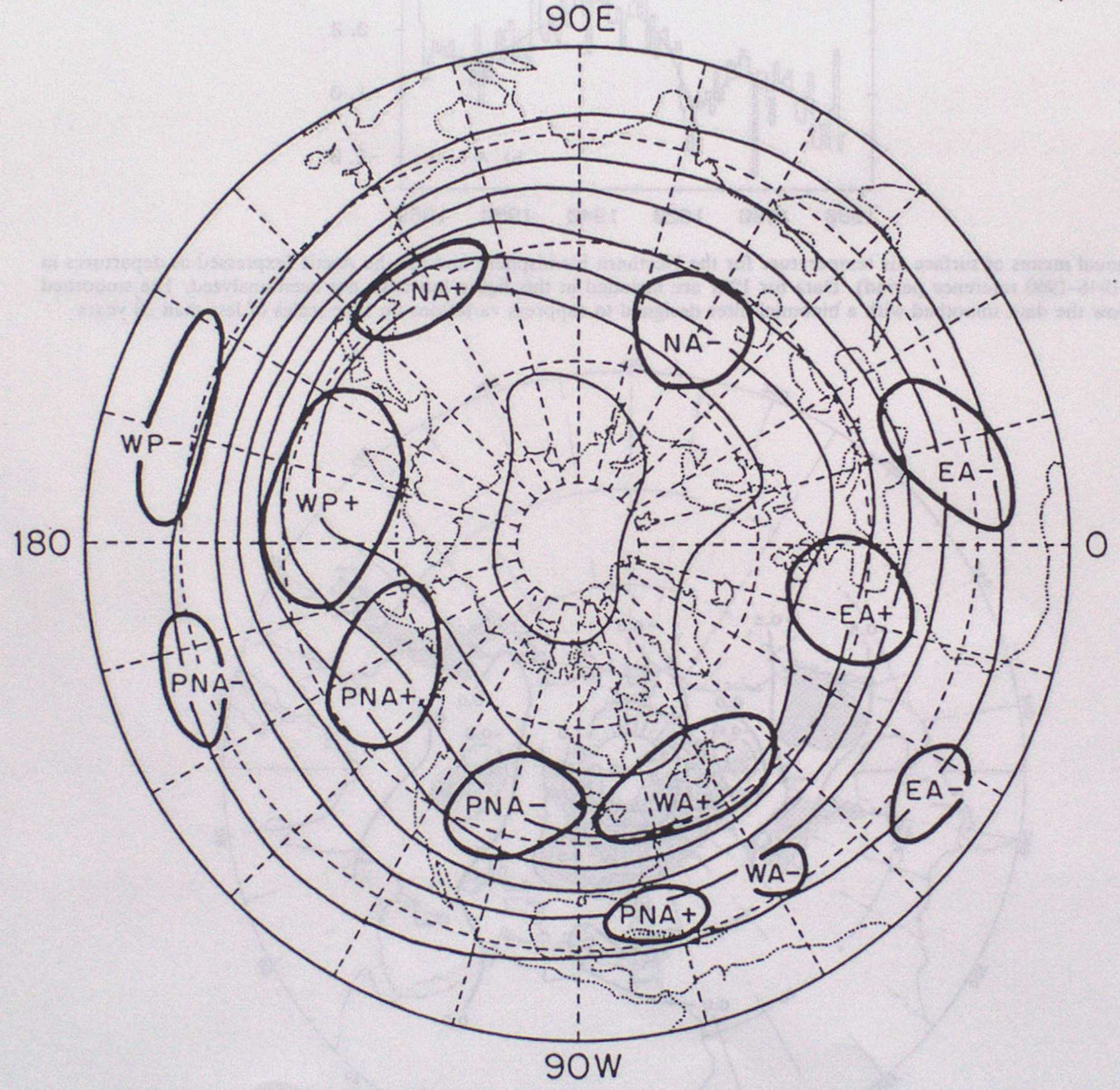


Figure 1.7 Teleconnections in Oregon State University AGCM also seen in a similar way in the atmosphere in northern winter. After Qiu and Esbensen (1984).
The "centres of action" of the teleconnections in the simulated 400mb height field are shown. The isopleths of the ±0.6 correlation coefficient between each of the five pattern indices and the 400mb height anomaly field are indicated by heavy lines.
"PNA" is the Pacific North American Pattern
"WP" is the West Pacific Pattern
"EA" is the East Atlantic Pattern
"WA" is the West Atlantic Pattern
"NA" is the North Asian Pattern

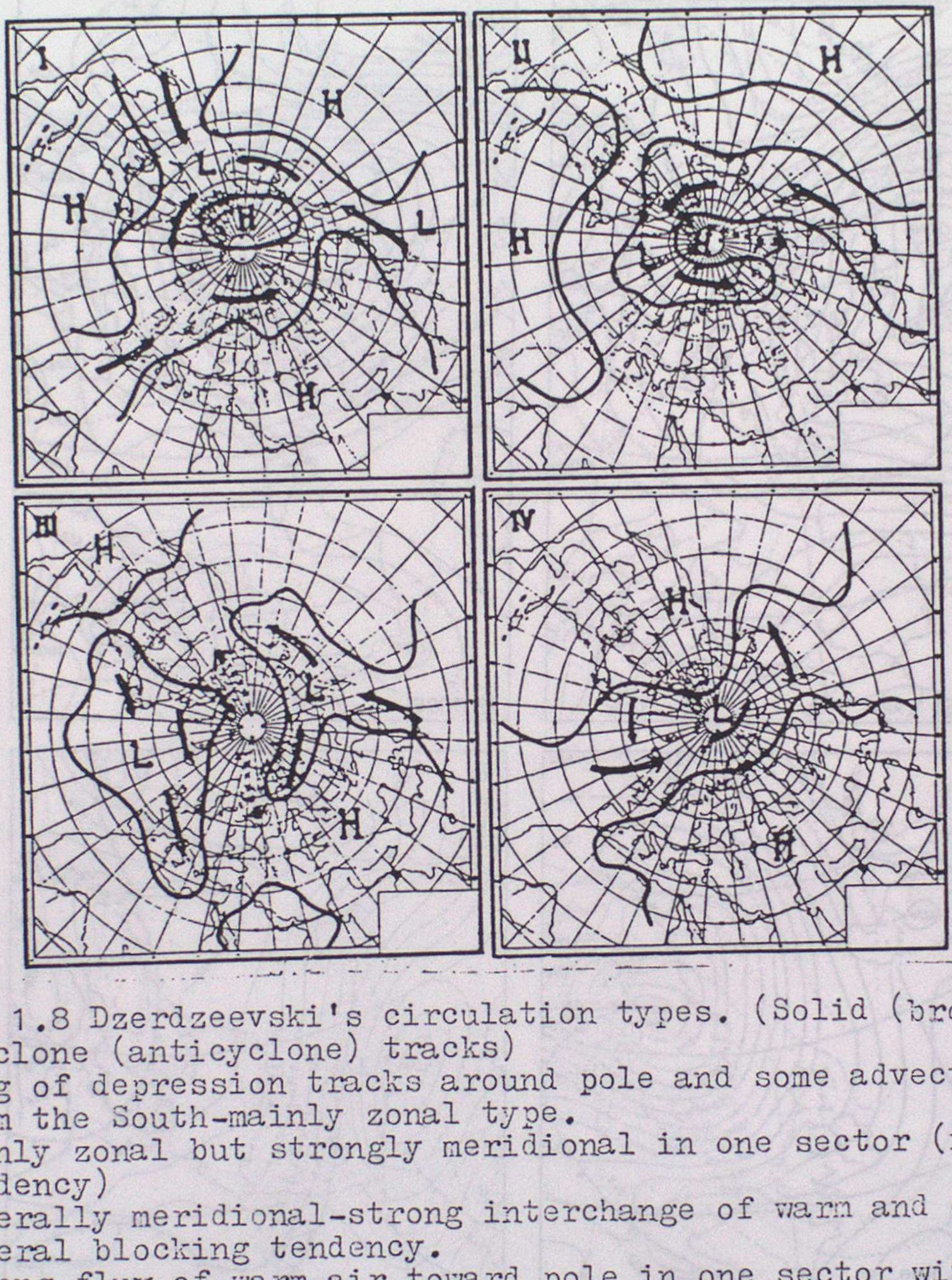


Figure 1.8 Dzerdzeevski's circulation types. (Solid (broken) arrows are cyclone (anticyclone) tracks)

I. Ring of depression tracks around pole and some advection from the South-mainly zonal type.

II. Mainly zonal but strongly meridional in one sector (ie blocking tendency)

III. Generally meridional-strong interchange of warm and cold air-general blocking tendency.

IV. Strong flux of warm air toward pole in one sector with depressions moving near central Arctic (Atlantic sector favoured in Winter).

After Barry and Perry (1973)

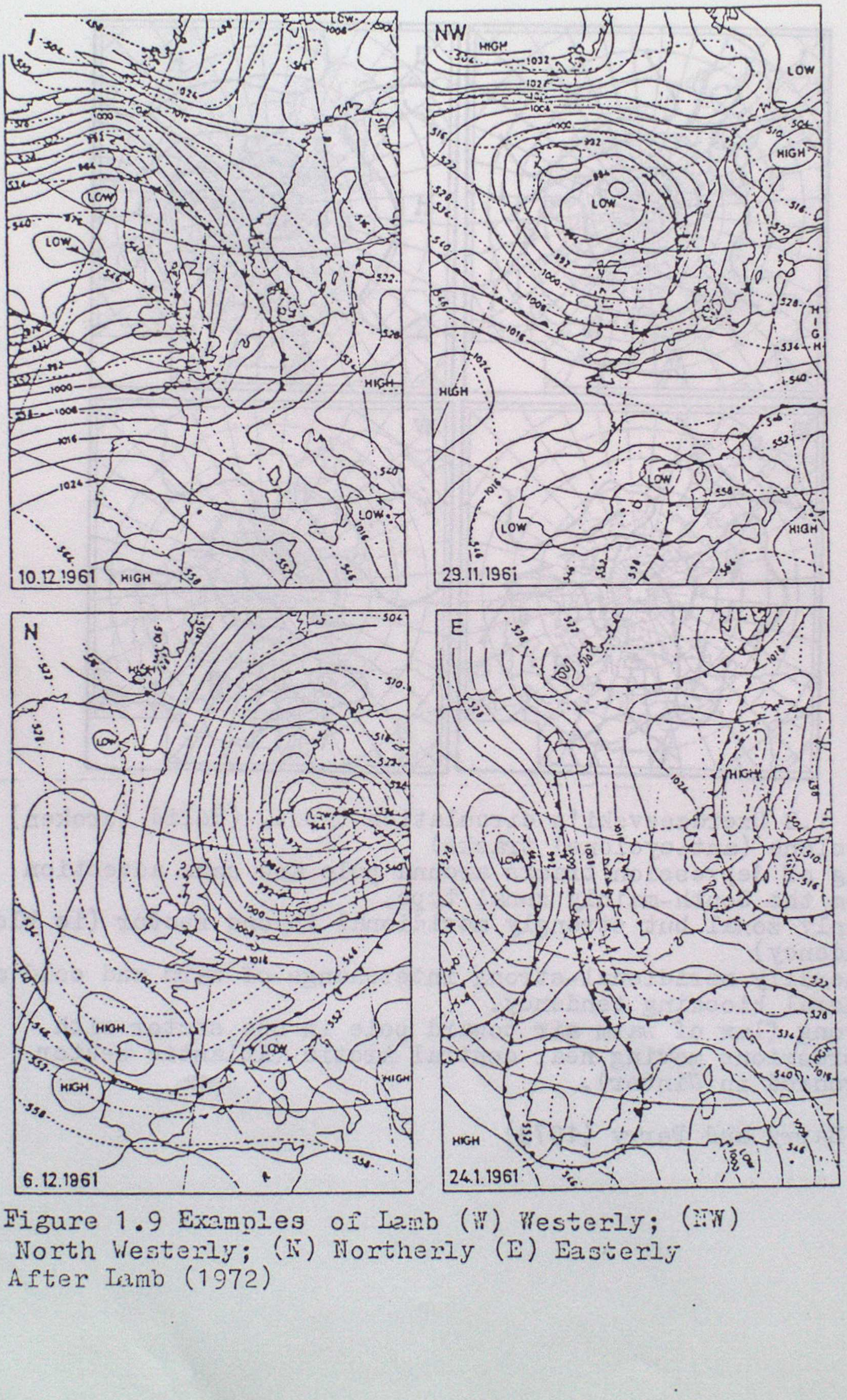


Figure 1.9 Examples of Lamb (W) Westerly; (NW) North Westerly; (N) Northerly (E) Easterly
After Lamb (1972)

ADVANCED LECTURE NO 2

DESCRIPTION OF WORLD CLIMATE

2.1 Introduction

The lectures on General Circulation (GC) have shown that ideas of a simple symmetric atmospheric circulation are far from correct but nevertheless useful. Fig 2.1A shows an idealised wind circulation for the globe while Fig 2.1B gives a more realistic picture where the central role of the mid-latitude transient eddies is shown. Fig 2.2 is a representation of the N Hemisphere mean meridional circulation: the strong upward motion at the ITCZ (whose position is variable) is shown, helping to give rise to a Hadley cell to the north with descending motion around 30°N (further north in summer). A "Ferrel" cell is shown further north still: it is the net result of eddy activity in mid-latitudes on several time-scales. Finally there is a more doubtful Polar cell.

Even in the Tropics, a simple picture of dominant mean motion resulting from a Hadley cell is rather misleading: besides eddy motions, there are a substantial set of more or less east-west circulations whose east-west components are called "Walker" cells (Fig 2.3). These cells are thought to be driven by the strong longitudinal variations of thermal forcing near the equator. The most prominent cell has a rising branch over and near the Tropical W Pacific in northern winter extending over SE Asia in the northern summer. The descending branch occurs over the relatively cool tropical east Pacific just to the west of S America. Fig 2.4 shows, from Newell et al (1974), estimates of the mean mass flux in the Walker circulation cells: the strongest mass fluxes are only a little less than those in the Hadley cells in Spring and Autumn, (Newell et al (1969)).

A partial explanation of these cells, which affect precipitation in equatorial regions very strongly, lies in the geography of tropical continents and oceans. Fig 2.5 shows the January and July climatological mean SST (for 1951-80) derived from the Meteorological Office Historical Sea Surface Temperature data set (MOHSST) (Minhinick and Folland (1984)). In the tropical Pacific and the tropical Atlantic there is a west-east surface temperature gradient. In the Pacific, the difference in SST at 10°S between 150°E and the S American coast is about 6-7°C in January (southern summer) and almost 10°C in July. In the Pacific, these longitudinal SST differences are comparable to those between the Equator and nearly 30°S in the same seasons, which of course help to drive the regional Hadley circulation. Not surprisingly, convective activity is usually very weak or non-existent in the tropical east Pacific south of the Equator but this situation is apparently rather unstable and a dramatic warming of the east tropical Pacific can sometimes occur (the El Nino-Southern Oscillation (ENSO) events, lectures 4 and 5). The result of these mean motions is that air-flow trajectories in the tropics are far different from those that could be derived from the simple picture of a meridional Hadley cell superimposed on an otherwise rather zonal circulation.

Fig 2.6 shows a map of global ocean currents which helps to explain the SST variations. The currents are of course driven mainly by the pattern of surface winds (Fig 2.7). The tropical anticyclonic ocean gyres are driven by the trade-wind circulations to the south and the westerlies to the north. These gyres are associated with strong, warm western ocean boundary currents: the Gulf Stream (GS) and Kuro Shio (KS) in the N Hemisphere, and in the S Hemisphere the Brazil (BC) and Agulhas (AG) currents (the latter off S Africa). Eastern boundary currents are cold: the Canary (west of N Africa) and Californian (CC) currents in the N Hemisphere and the very cold Humboldt (HC) and Benguela (BC) currents in the S Hemisphere. So it is not surprising that SST is very low, for example, off Peru though upwelling of water from below is an additional factor in all cold currents in tropical regions. Upwelling results from an off-shore Ekman drift which is driven by the prevailing off-shore winds. In the Indian Ocean, currents tend to reverse in direction with the seasons and are driven by the SW (Summer) and NE (Winter) monsoons.

2.2 Seasonal variations of atmospheric circulation

The climate of the world can only be understood by including the seasonal variations in atmospheric circulations, snow and ice etc.

2.2.1 January and July pressure and wind fields

Important features in Fig 2.7-2.8 are:-

January - Note the subtropical highs; the low pressure trough of the ITCZ near to or south of the Equator; the intense Siberian high with surface air streaming away from it. Being a cold high, the Siberian high decreases in intensity with height and is overlain by the midlatitude westerlies. In the N Pacific and N Atlantic there are deep lows (the Aleutian and Iceland lows).

In the S Hemisphere (Summer) the subtropical highs form an almost continuous belt, with centres over the cool oceans; heat lows dominate the three main continental land masses.

July - The northern subtropical highs are further north with considerable PMSL differences between land and sea in the same latitudes; heat lows exist over the southern United States, the Sahara, and most prominently, southern Asia. The Iceland low is still present but rather weak but there is now very little sign of the Aleutian low, the east Pacific high being placed surprisingly far to the north. Fig 2.7 shows that the ITCZ moves further away from the Equator over land than over the oceans. Being largely the result of vertical convective processes, the tropical lows shown in Fig 2.8 decrease in intensity with height unlike the Iceland low, which merely slopes to the west with height.

Note that the ITCZ lies where the trade winds meet. When this is far from the Equator, the air that crosses the Equator turns southwesterly in direction due to the change in direction of the Coriolis force though a westerly component sometimes develops before the Equator is crossed, eg over the Indian Ocean in July. Fig 2.9 shows that the picture of a single ITCZ moving North over the Indian Ocean in summer is far from correct. The apparent double ITCZ structure, if that is what it is, is poorly understood but is the subject of current research. A possible explanation is that individual zones of tropical convergence tend to occur above individual absolute SST maxima if those maxima exceed about 27-28°C (Gadgil et al

(1984)). On Fig 2.5 for mean July SST, there is a suggestion of a SST maximum near the southern of the two convergence zones as well as a further SST maximum adjacent to the coast of the Indian sub-continent.

2.2.2 Upper level flow

2.2.2.1 Tropics and Sub-tropics

Figs 2.10 and 2.11 show the 850 mb and 200 mb streamlines during Dec-Feb and June-Aug. At 850 mb, the anticyclonic centres over the oceans are clearly seen in both seasons, overlain by a strong westerly flow at 200 mb. In June-Aug a strong warm anticyclonic centre is seen at 200 mb over N India/Tibet, with cyclonic motion at 850 mb. This vertical structure is the result of convective heating over N India and heating of the very elevated Tibetan plateau (height 4000-5000 m) resulting in a high-level warm anticyclone over this region; this gives rise to a strong 200 mb easterly flow to its south which extends all the way from the tropical W Pacific to the tropical Atlantic. At the surface, the flow is mainly westerly. Over coastal W Africa Fig 2.7B indicates a south westerly flow at the surface though at 850 mb Fig 2.11 suggests a more south to south-easterly flow. A problem in this region (see later) is that interannual and longer time scale variability has been very large in recent decades so different, rather short, data periods may give different results. In the S hemisphere in Dec-Feb, a monsoon-like heat low over S Africa at the surface (cyclonic centre marked C in Fig 2.10) is overlain by a warm 200 mb anticyclonic centre; there is a similar feature over the Bolivian Plateau. To the north and north east of the latter feature, there exists strong mean 850 mb north westerly flow over W Brazil and W Argentina which is evident also in June-Aug. This north-westerly tropical circulation is thought to depend on the fact that the Andes form a sharp physical western barrier to flow over these regions (Schwerdtfeger (1976)). This is the only tropical region in the S Hemisphere where low level tropical north westerlies are common (except perhaps near Papua-New Guinea in northern winter).

2.2.2.2 Mid-latitudes (500 mb)

Figs 2.12A,B from Palmen and Newton (1969) show the 500 mb flow in the N Hemisphere with its attendant ridges and troughs. Here it is worth mentioning that the S Hemisphere 500 mb flow is more symmetrical but there is a marked tendency to a broad trough over the Southern Ocean between about 30°W and 40°E. This feature is associated with long-term SST and air temperature averages up to 5°C below the average for their latitude between 50°-60°S (Hoflich, (1984)). Fig 2.13 shows the mean S Hemisphere 500 mb flow in January (summer). Between about 40°S and the S Pole, mean 500 mb flow is of similar strength to that in the N Hemisphere winter in a similar latitude band (which of course includes the same month, January). In the S Hemisphere the strength of the flow in local winter hardly changes from that in summer because the surface temperature gradient between equator and pole scarcely alters through the seasons.

2.3 Mean seasonal snow and ice variations

Ice and snow are an important feature of the mean seasonal variation of climate (Fig 2.14A). Taken together, the regions of permanent sea and land ice in the N Hemisphere are currently slightly smaller than those in the S Hemisphere, but the maximum seasonal extent of ice and snow in winter is much larger in the N Hemisphere. Sea ice around Antarctica fluctuates from a typical minimum extent of 2.5×10^6 km^2 (Feb/March) to a typical maximum of 15×10^6 km^2 (September) (note that the area of Antarctica itself is 14×10^6 km^2). Over the N Hemisphere the typical minimum extent of sea ice is about 8×10^6 km^2 (September) with an additional 1.8×10^6 km^2 of land ice contributed by Greenland. However the maximum extent of ice and snow together is apparently about 55×10^6 km^2 in the N Hemisphere (February) (Untersteiner (1984)). There are pronounced interannual fluctuations of N Hemisphere snow and ice extent (see Fig 2.14B from Hahn (1981)). Reliable measurements of fluctuations in Antarctic sea ice on time scales of several years are just beginning to emerge (Chiu (1983)). Reliable satellite plus conventional assessments of global seasonal cryospheric extent only began about 1974.

2.4 Climate classification and world climate

In one lecture I cannot attempt to cover the climate of the whole world in detail. I shall briefly discuss climate classification and then select a few main regions for discussion. I shall also mention some lesser known but perhaps thought-provoking habits of regional climates.

2.4.1 Climate classification

The climate of even a small area is composed of a large number of locally-varying climate elements and it is nearly impossible for two places to have identical climates. So condensation and ordering is needed to facilitate analysis and explanation. Classification consists of recognising important characteristics which individuals have in common and grouping the individuals into a few classes or types in order to establish generalisations.

Classifications can be based on the way climates are generated or on their effects eg their effects on surface vegetation or specialised practical problems like the reliability with which maize can be grown in UK discussed in Lecture 1. In fact applied classifications are now numerous but are usually of main interest to specific users, though as a result many useful climatic maps of extremes or the frequencies of given thresholds of rainfall amount, temperature, wind speed etc have been produced.

The genetic type of classification appeals to meteorologists but it must be simple and there is a danger that the facts of climate may be forced into a preconceived and not wholly appropriate explanatory structure. Empirical classification starts from the observed facts but the basis of division can be arbitrary and its validity depends of course on the use to which it is put.

The best known empirical classification is that due to W Koppen (1931). A considerable number of Koppen's climatic types and subtypes do in fact coincide with broadscale features of the general circulation which allow the numerous and important deviations from a "standard" world pattern to be realistically highlighted. Koppen's system is known throughout the world and was the result of half a century's study and refinement. His classification is quantitative and is based upon annual and monthly means of temperature and precipitation; it accepts native vegetation as the best expression of climate so that many of the climate boundaries are selected with vegetation limits in mind. It recognises that plant development depends on available water as well as on temperature so that precipitation and evaporation are included in the classification scheme. Because evaporation measurements are not usually available, evaporation is estimated from temperature and precipitation. Many authors have proposed modified schemes based on Koppen's scheme; they mostly differ in the method of expressing the available water for plant growth. Appendix 1 to this lecture describes Koppen's scheme in detail and has been taken from Lamb (1972). It is interesting that Koppen's scheme has already been used to help interpret the results of an AGCM experiment (Manabe and Holloway (1975)).

Fig 2.15 shows a climate classification scheme due to Critchfield (1983) which is rather like a simplified version of Koppen's or Trewartha's system, and we shall use Critchfield's scheme as a background for our description of world climate. Three main climatic classes are defined (subjectively, not rigorously) in terms of monthly mean temperatures (basically according to freedom from frost and the length of the growing season) with an extra highland climate class which includes all temperature classes and can occur anywhere in the world. Arid climates are a subgroup of the three main types where precipitation is less than potential evaporation. Like most schemes, Critchfield's scheme subdivides classes according to whether or not there is a dry season and when in the year it occurs (except for highland climates). Fig 2.16 shows a map of mean annual world precipitation which provides a useful background to the following description.

2.5 Description of selected world climates

2.5.1 Tropical climates

Consider first the air flowing around the oceanic cells of the sub-tropical high pressure belt centred about 25-35°N and S. The air flowing toward the equator on the eastern side of the cells and not far from the latitude of the centre of the cells is stable and dry and strongly subsides due to the large-scale divergence caused by the Hadley cell. This gives rise to a low level temperature inversion (Riehl (1979)) (Fig 2.17). As a result, the eastern half of the sub-tropical oceans has a sunny and fairly dry climate so that islands such as the Canaries tend to be arid, though orographic precipitation on high mountains can be considerable. The subsidence also gives rise to extremely arid coastal regions on the western sides of the continents in latitudes 20-30°N and S such as the Kalahari and Chile-Peruvian deserts and the western margin of the Sahara.

The dry Equator-ward moving air picks up heat and moisture from the sea, so that the trade-wind inversion steadily rises as the air turns more and more to the west. The air beneath the Trade-wind inversion becomes deeper and more unstable and increasingly subject to convection. Riehl shows theoretically and from observations that the trade wind inversion is not a physical barrier: sinking and diverging air from above descends through the inversion as shown in Fig 2.17, the descending motion being forced by the large scale divergence around the high pressure cell. The descending air is moistened as it sinks; thus the inversion rises as more air is incorporated beneath it. The air becomes quite rapidly warmer and moister because sea temperatures also tend to rise towards the western halves of the tropical oceans.

As the air approaches the Equatorial trough of low pressure, it meets the corresponding flow from the other hemisphere and is subjected to strong convergence and lifting, and so creates the great deep convective cloud systems of the ITCZ. Although the Trade Winds are mostly dry, fair-weather winds on the eastward and poleward sides of the high pressure cells, their Equatorward and western sectors are subject to synoptic-scale disturbances including tropical storms in the western sectors if SST is high enough and the latitude exceeds about 8°. So the amounts of cloud and precipitation in the west are considerably greater though most rain falls near the ITCZ itself. Much of the latent heat energy released in the upward branch of the Hadley cell re-enters the lower atmosphere well away from the Equator and a considerable amount of this energy is conveyed back towards the Equator by the trade winds (thus the trade winds tend to "heat the equator"). Upward motion in the region of the ITCZ is confined at any one time to small areas concentrated in mesoscale convective systems which have a bewildering range of sizes which are poorly understood. The ITCZ follows the annual path of the vertical midday sun but moves much further in some longitudes than in others, tending to move most over continents. So a fundamental division of tropical humid climates is between those regions near the Equator which have considerable rainfall all the year round (tropical wet climates or equatorial rain climates) and those regions which have a pronounced dry season in the local winter when the ITCZ has moved well away or even into the other hemisphere so that the region is subject to subsidence from the sub-tropical anticyclone (tropical wet and dry climates).

This difference of climate has a great influence on vegetation and marks the difference between the jungles, the hardwood forests and swamps of the Congo or the Amazon basin, south east Asia and parts of Central America and the lighter more deciduous forests and tree studded grasslands often called savannah which are typically located to the north and south of the equatorial region. Good examples of savannah exist in South Central Brazil, northernmost South America and in tropical Africa. It is not certain whether the existence of savannah is always caused by drought conditions in winter or whether it is caused by the frequent fires that sweep the grasslands from which trees do not recover. It is interesting that man himself in recent centuries may have inadvertently created apparently natural savannah. New savannahs are now being deliberately created, especially around the edges of the Amazon jungle.

Much work on the General Circulation and climate concentrates on the tropical and equatorial zone: this is the major source of energy for the atmosphere and it covers about 40% of the world. It is also one of the zones about which least is understood. It was once thought that tropical climates were simple and followed regular patterns but this is now known to be far from correct. Most importantly, the physical details of the climates are not easily explained, especially their tendency to pronounced inter-annual variability.

Africa probably provides the best examples of the expected arrangement of climate zones of the tropics so we shall look at the structure of the atmosphere over West Africa where the arrangement of climates zones is simplest. Please refer to Figs 2.7, 2.8, 2.10 and 2.19 and note the following:-

a. The S Hemisphere south east Trade Winds cross the Equator to become moist, south westerlies (sometimes called "monsoon" air). (Fig 2.7B).

b. The north east Trade Winds over the Sahara originate from near the Mediterranean (and their specific humidity is fairly low). (Fig 2.7B).

c. The ITCZ does not normally move south of the Equator in the longitude of W Africa even in winter because the tropical S Atlantic Ocean is relatively cool. This means that the Gulf of Guinea coastal region is wet in the winter.

d. Further east in tropical Africa, the behaviour of the ITCZ becomes more complex especially in Ethiopia, Kenya and Uganda. It is much affected by the high mountains and plateaux of this region and in summer by the Indian monsoon circulation. So this is a poorly understood region - one of the earth's "problem climates".

The vertical structure of the atmospheric flow in the tropical W African region in August and some of its interannual variability is shown as a function of latitude in Fig 2.19A and B (Newell and Kidson (1984)). These diagrams show the zonal and meridional components of the wind through the depth of the atmosphere for two sets of recent years. In one set the Sahel region (Region 5 south of the Sahara desert in Fig 2.15) was wetter than normal and in the other set it was dryer than normal. Referring to Fig 2.11, we would expect to see, north of about 5°S, fairly strong upper level easterlies extending from the Asian monsoon circulation. These easterlies are clearly present in both sets. In the years marked wet, the upper easterlies are slightly stronger than in the years marked dry but the sample is too small to be sure of this. Lower down, there is a pronounced easterly jet near 700-800 mb near 20-25°N which is more pronounced in the dry years. This result is probably real and due to an enhanced low-level thermal gradient between the wet regions near the coast and the more than usually dry region of the North Sahel. This dryness should give rise to more sensible heating of the ground and thus should warm the lower layers of the overlying atmosphere. A low-level south westerly flow is evident in both wet and dry years between 10°S and 20°N but is markedly less deep and certainly is less strong in the dry years. In fact the ITCZ annually reaches 20-22°N where practically no rain falls; the boundary between the northerly and southerly components of the flow near the ground in August does not seem to vary greatly between wet and dry years. However Fig 2.19B suggests that the flux of warm, moist air from the S Atlantic is weaker in the dry years.

The detailed mechanisms of rainfall release near the ITCZ are complex and ill understood but certainly are not solely accomplished by the random occurrence of isolated thunderstorms. Fig 2.20, taken from Riehl (1979), shows a typical organised synoptic disturbance, known as an "easterly wave", which was observed during GATE (The GARP Atlantic Tropical Experiment of 1974). When viewed from a satellite, easterly waves can look like a very large mesoscale "cloud cluster" though there are other types of cloud cluster. In Fig 2.20 the winds are south west to the south of the trough line and north easterly to the north so diagrams of seasonal mean conditions like Fig 2.19A and B are the net result of many individual synoptic features.

Referring back to Fig 2.15, we can see several nearly parallel climatic zones in W Africa. Zone "4" is the western Sahara desert where rainfall is always rare; zone "5" is the Sahel which is placed north of about 10°N in the west of W Africa but which tends to be located a few degrees to the south further east. The Sahel depends, as discussed above, for its rainfall on the northward movement of the ITCZ into the region in July-Sep. Area "3" is fairly wet most of the year as it is affected by the ITCZ, or the moist unstable air just to its south, much of the time. The small area "5" in south Togo is sufficiently far enough south in summer to escape much influence from the ITCZ so has a summer dry season. The very wet region "2" around Liberia and Sierra Leone is an anomaly probably partly caused by local convergence due to mountains just inland from the coast, otherwise it would be like area "3".

Atypical of these latitudes is the dry zone of tropical East Africa near the Indian Ocean coast that stretches from about 4°S to at least 10°N. This region has been the subject of much study. The Indian (northern summer) monsoon flow strongly affects this region and the moist easterlies experienced by other continental eastern margins near the equator are replaced by a south-south westerly flow that seems to be relatively dry but does not extend very far inland. The lack of rain in the south Somalia/Kenya desert region in southern summer seems to be related to an airstream which is derived from an anticyclone over Arabia rather than from the warm ocean but the lack of rain in northern Somalia in the northern summer does not have an adequate explanation. Perhaps the influence of strong seasonal upwelling of cold water offshore near 10°N during the Indian south westerly monsoon season plays a part in the dryness, in addition to that of the dry southerly airflow.

I shall not discuss the Indian or other monsoon in more detail: please see Riddaway (1980). In passing one might note that much of India is classed by Critchfield (1983) (Fig 2.15) as a tropical wet and dry climate though the dry season on much of the west, mountainous, coast is short and there is enough soil moisture during the dry months to support tropical rain forests. The true tropical wet climate starts further east in Burma. A typical example of this climate is enjoyed by Belem at the mouth of the Amazon. The mean temperature of every month is close to 26°C while average monthly rainfall varies between about 100 and 450 mm. Sunshine averages 5.5-6 hrs a day. So even here a short dry season exists and in fact only restricted areas of the equatorial regions have a uniformly wet climate eg in the western Amazon basin and in western Colombia on the other side of the Andes north of the Equator. This remarkable climate, which is hardly understood, is uniformly extremely wet; Fig 2.21 shows daily rainfall for the year 1928 which totalled 7089 mm and was remarkably uniformly distributed through the year. By contrast there is a small, heavily populated, dry area in NE Brazil (the "Nordeste") (Fig 2.22) whose climate is again ill-understood but which is being intensively studied (Hastenrath et al (1984)). Much hardship is quite regularly caused here by drought.

Turning to an ocean climate, the "desert islands" just south of the Equator in the central and eastern Pacific stretching from near the dateline to the Galapagos Islands represent one of the most extraordinary of world climates. These islands stretch a quarter of the way round the world and owe their unusual dryness to the fact that a cool sea is normally maintained by the wind circulation so that the ITCZ is almost always north of the Equator. The cool sea near the Equator seems to be maintained by Ekman divergence. However, every few years the sea becomes very warm and the ITCZ moves over the area for much of the high-sun season, giving intense convective activity. Indeed the tropical sea surface temperature maximum, viewed in the meridional direction tends to move over the islands. Average yearly rainfall totals of 500-700 mm are quoted for many of the islands but this hides the fact that in dry years perhaps only 50-150 mm fall and in the few wet years there may be as much as 2500 mm. These fluctuations are now known to be intimately related to the ENSO phenomenon which happens sufficiently often that these islands could be truly said to have a bi-modal "almost intransitive" non-seasonal climate (see Fig 1.4 in Lecture 1, and Lecture 5).

Before leaving the tropics, we might note some interesting features of highland tropical climates. Tropical glaciers are surprisingly numerous eg there are many in the tropical Andes, several in Africa in the Ruwenzori Mountains of Uganda and on Mounts Kenya and Kilimanjaro above about 4500 m. Surprisingly there is even a 7 km^2 ice cap in the Star Mountains of W Irian, Indonesia, known as the Caarstenz Glacier (McAlpine et al (1983)). This glacier exists at a height of about 4500-5000 m despite being surrounded by the warmest SST in the world. However at about 15°-25°S in the Andes no glaciers are found at these or even much higher levels. The determining factor here is lack of precipitation and lack of cloudiness so only very small glaciers are found above 6000 m. By contrast in the cloudy Andes of Equador near 0° one glacier extends down to almost 4300 m. Many tropical glaciers are slowly receding and may not be in equilibrium with the current tropical climate.

2.5.2 Subtropical climates

2.5.2.1 Mediterranean climates

On the western margins of the continents in both hemispheres there are regions where summer is dominated by the anticyclones of the desert zone (ie the descending branch of the Hadley circulation); these anticyclones are particularly stable being on the eastern cool sides of the subtropical oceanic circulation. However in winter these areas come under the influence of mid-latitude travelling cyclones and are consequently much wetter with changeable weather. These are the areas of Mediterranean climate, characterised by (1) a concentration of the year's rainfall in winter and sometimes in autumn, while the summers are mainly dry, (2) they have warm to hot summers and mild winters, (3) there is abundant sunshine especially in summer. The five regions where the Mediterranean climate occurs are fairly similar in latitude: the border lands of the Mediterranean Sea stretching for 2500 km into Eurasia, the coastal strip of central and southern California, central Chile, the south western tip of Africa and parts of southernmost W Australia, S Australia and Victoria.

In California and Chile the Mediterranean climate ends abruptly at high mountains 100-300 km inland with mountain-induced desert on the eastern side. The remaining continents of the S Hemisphere only just reach the appropriate latitudes while the Mediterranean Sea itself acts as a cool-season convergence zone and route for winter depressions. In this respect the Mediterranean is different from the other areas and some of these winter depressions penetrate well into Asia.

It might be thought that Mediterranean summer climates would always be hot but in coastal S California and Central Chile cold offshore currents (see Fig 2.5 and 2.6) keep coasts 10°C or more cooler than inland and more than this by day. The coasts have frequent fogs and low stratus and strong sea breezes set in because of the contrast in afternoon temperature with the hot interior. In the poleward parts of these regions occasional winter frosts occur and as the winter of 1984-5 showed on the French Riviera, snow may on rare occasions reach down to the coast. (Eastern Spanish beaches in January 1963).

2.5.2.2 Subtropical Eastern sides of continents

The subtropical eastern sides of continents receive, in summer, unstable or neutral tropical maritime air because subsidence in the subtropical anticyclone on the western side of the oceans is weak and the atmospheric boundary layer is moist. This favours convective activity and abundant summer rain aided by the development of tropical storms, some of which gain the structure of mid-latitude depressions as they pass over or near these regions. There is also a tendency for monsoon circulations to develop, especially in south east China and to some extent in the south east United States. This climate is also represented in all 3 southern continents, ie S Brazil, East Australia and SE South Africa. It differs from the Mediterranean type by having tropical humid climates on its equatorial flank, instead of tropical hot dry climates. These tropical humid regions provide a source of sultry heat in summer rather than the dry heat that exists adjacent to the Mediterranean climates.

In the N Hemisphere, a severe continental climate exists on the poleward flank of the subtropical eastern continental climates which can bring intense cold in winter for brief spells. For example, Nanking in eastern China at 31°N has occasional severe snow storms and frequent frosts in winter. Even further south, in N America, occasional very low night minima are a feature of the climate of central and northern Florida. Summers are hot, the monthly mean temperature averaging 24-27°C or slightly higher, with high accompanying humidity from the dominating tropical maritime air, giving rather oppressive nights. Annual rainfall is relatively high (1000-1500 mm is typical) and tends to be concentrated towards the summer and autumn seasons, when tropical cyclones are most frequent and when cold fronts moving from the north-west can give very heavy precipitation, especially in USA.

The subtropical area of south east Asia has a complicated, ill-understood, climate. In winter the subtropical branch of the jetstream flows around the southern slopes of the Himalayas, while the polar-front jet flows to the north of the Tibetan plateau; both meet in an upper convergence zone over China. This upper convergence zone leads to subsidence over inner China and therefore to generally fine weather in winter. The two jet streams tend to reinforce each other over Japan, forming part of the very strong E Asian trough and giving extremely high wind speeds which can regularly reach 100 m/sec at the 200 mb level. In summer the subtropical jet stream moves to the north of Tibet and the polar front jet recedes further north so that the upper convergence zone over China disappears. In Japan an early summer rainfall maximum occurs, the "Bai-U" rains, followed by a later summer maximum. Fig 2.23 gives a schematic picture of the surface wind fields over SE Asia in summer (SW monsoon) and in winter (SE monsoon).

2.5.3 Mid-latitude climates

2.5.3.1 General Discussion These are dominated by the influence of the upper westerlies and their attendant troughs and ridges and on the western sides of the continents the surface westerlies also very prominent. So the natural subdivision is into maritime and continental climates. The maritime climates are found on the extreme western sides of N America, over much of Europe, and in the S Hemisphere in Southern Chile, in Tasmania and over most of New Zealand. Truly continental climates are found (in the N Hemisphere) either well inland or on the eastern sides of the continents where extreme seasonal variations of temperature result from their leeward location and the influence of summer monsoonal wind systems. Patagonia in Southern Argentina has some of the characteristics of an eastern mid-latitude climate but the continent is too narrow for the temperatures to be extreme while the influence of the Andes renders the climate extremely dry.

The mid-latitude oceanic climates mostly lie poleward of 40° and stretch well towards the poles because of the influence of ocean currents which are warm or relatively warm for their latitude. In N and S America the climates occupy very narrow belts only 200-300 km wide with an extraordinarily rapid transition zone to dry climates to the east. For example in parts of Northern Oregon and S Washington states the transition from the mid-latitude oceanic to the inland dry climate can be accomplished in a day's walk. The major characteristics of this oceanic climate are cool summers, mild winters, changeable weather, considerable cloudiness, and mostly adequate rainfall at all seasons. The seasonal emphasis in rainfall tends to be towards late autumn and winter on the extreme western margins, and late summer and early autumn further inland. Over north west Europe, there is a tendency for late spring to be driest, due to a peak frequency of blocking anticyclones at that season. Nevertheless seasonal rainfall deficiency does not cause a dormant period for plant growth.

Unusual aspects of the mid-latitude oceanic climate include the rather dry summers in the north west United States, probably caused by the northerly extension of the N Pacific high over the Gulf of Alaska, and the exceptionally mild winters over large areas of W Europe and the coastal fringe of Norway beyond the Arctic Circle, caused by the openness of these regions to the warmth of the eastern N Atlantic. The oceanic margins also exhibit a strong decrease of temperature with altitude and therefore in the length of the summer growing season. The tree line in NW Scotland is at only about 1000 ft, though man has removed some of the remaining trees and it sometimes appears to be almost at sea level. This very low tree line is also partly a result of the exceptional windiness of the climate.

The Chilean mid-latitude climate is one of the cloudiest in the world as frequent depressions in the stormy sub-Antarctic summers and winters (see Fig 2.13) meet the formidable, steep, highland barrier of the southern Andes. The combination of extreme cloudiness, heavy precipitation and steep mountains has created one of the most remarkable climatically-caused phenomena in the world. Fig 2.24 and Charles Darwin's words (1979 reprint) speak eloquently:- "In Eyre's Sound, in the latitude of Paris, there are immense glaciers, and yet the loftiest neighbouring mountain is only 6200 feet high. In this Sound, about fifty icebergs were seen at one time floating outwards, and one of them must have been at least 168 feet in total height. Some of the icebergs were loaded with blocks of no inconsiderable size, of granite and other rocks, different from the clay-slate of the surrounding mountains. The glacier furthest from the Pole, surveyed during the voyages of the Adventure and Beagle, is in lat 46° 50', in the Gulf of Penas. It is 15 miles long, and in one part 7 broad, and descends to the sea-coast. But even a few miles northward of this glacier, in the Laguna de San Rafael, some Spanish missionaries encountered "many icebergs, some great, some small, and others middle-sized," in a narrow arm of the sea, on the 22nd of the month corresponding with our June, and in a latitude corresponding with that of the Lake of Geneva!"

These glaciers, descending from the vast mid-latitude Patagonian ice-sheet, set Darwin thinking about the causes of the ice ages in subsequent pages of "Voyage of the Beagle".

The mid-latitude temperate continental climates (here defined as having a cold month under 0°C) exhibit a cold winter, with a durable snow cover, and a warm summer season, so that the annual range of temperature is large. In the maritime climate the temperature of one winter is much like that of another, but wide deviations from the normal seasonal temperature are characteristic of continental mid-latitude climates (as much as 10°C in extreme climates). Summer is the season of maximum rain, which is mostly convective.

A feature of some northern hemisphere mid-latitude climates is "blocking" which is a strong influence on UK weather variations. Fig 2.25 from Lejenas and Okland (1983) shows the longitudinal variation of blocking in the N Hemisphere. In the S Hemisphere blocking is rarer but not uncommon near New Zealand for instance. The effect of blocking on weather over a month or a season can be surprisingly large. Fig 2.26 show two contrasting "blocked" and "westerly" Januaries over NW Europe (Folland (1983)). The reasons for blocking are currently an area of much research.

2.5.3.2 Continental anticyclones

Blocking is, superficially, a result of stationary large amplitude ridges in the westerlies: anticyclones associated with the block have a warm core. In winter, the Eurasian and N American continents have cold anticyclones as a prominent feature of the climate: they are situated to the west of the regional climatological mean upper troughs. These cold anticyclones do not extend very far upwards into the troposphere but bring intensely cold weather to the eastern half of the continents. The Siberian anticyclone is firmly sited well away from the western seaboard and is thus relatively undisturbed by mid-latitude depressions. However the N American winter anticyclone tends to be sited over or to the east of the Rocky Mountains and it is not as stable a feature as the Siberian high, often being disrupted by depressions moving in from the west so that the winter weather of Central and Northern USA is decidedly changeable. Surges in the Siberian anticyclone can bring intense cold to Japan and even to south China (a subtropical climate) so that even maritime Hong Kong has suffered air frost in winter (22°N).

2.5.4 Dry Climates

The essential feature of a dry climate is that the potential annual water loss through evapotranspiration at the earth's land surface exceeds the annual gain from precipitation. Thus dry climates do not have a constantly replenished ground water. Fig 2.15 shows that arid climates cover a broad latitude band and occupy a considerable part of the earth's land surface. The deserts are usually surrounded by semi-arid lands (called steppes in Asia) where rainfall is very unreliable.

2.5.4.1 Typical types of desert climate are:-

The tropical dry climates lying at about 20 or 25°N and S with their northern and southern margin at about 15° and near 35°. These climates coincide fairly closely with subsiding branch of the Hadley cell. The deserts do not reach the eastern coasts of continents (usually) because in these regions the subtropical anticyclones are much less stable. There are two types of semi-arid regions around these deserts: those that lie on their equatorial side and experience rainfall at the time of high sun due to the northward progression of the ITCZ, and those that lie on the poleward side of the deserts and receive rain from the mid-latitude westerly circulation in winter. In N Africa, parts of the Atlas Mountains and Tunisia can receive 500-800 mm of rain. Typical desert rainfalls are much less (Cairo 30 mm in a year; Yuma, Arizona 80 mm; Calama, northern Chile usually no rainfall in the year). Average summer temperatures are about 30-35°C so maxima are as a rule near 40-45°C, and in the winter months a mean of 10-15°C is typical. On poleward highland margins snow may fall and lie and mean winter temperatures may be similar to those of UK but with a much larger daily range. In the Central Asia deserts, winter temperatures are much lower (northern Tibet, China, parts of Mongolia).

2.5.4.2 Western ocean margins of dry climates

Here cool ocean currents strongly affect air temperature which can be as much as 10°C lower than normal for the latitude with a small annual range. There are surprising features of these dry oceanic climates, particularly their exceptional dryness eg in northern Chile. Lettau and Lettau (1978) has shown that along the western margin of the Chilean desert the juxtaposition of very cold sea and intensely heated land gives rise to a low level jet stream (500 m above the ground) which flows from south to north (Fig 2.27). This is considered to arise from a sea breeze type circulation which is so strong that it persists, less strongly, by night and thus all the time. If this circulation was spun up from rest as a "sea breeze" the Coriolis force would strongly influence it after a few days. Thus the wind tends to blow northwards parallel to the coast in geostrophic equilibrium with high pressure to the west and low pressure to the east but with a sea breeze circulation cell superimposed. The subsiding component of this sea breeze component occurs over a narrow band of ocean immediately adjacent to the coast and is thought to give rise to the long thin, cloudless strip of ocean off the Chilean coast often seen on satellite photographs. So air moving in this jet undergoes a corkscrewing, helical type of motion. The additional effects, if any, of the low level jet on rainfall are not clear.

2.5.5 Climate of the S Pacific and S Ocean 150E-70W

This is the region least well observed but it is large in extent, covering 120° of longitude, and has some unexpected features. Fig 2.28 shows that in addition to the expected features which can be deduced from Figs 2.8-2.11 and 2.13, there is a zone of cloudiness and cyclogenesis stretching from near the Equator across the whole width of the Pacific to mid-latitudes. Thus the ITCZ in the tropical W Pacific appears to be directly connected with the westerlies near Cape Horn. This "South Pacific Convergence Zone (SPCZ)" shows up as a cloud band which stretches south east across the S Pacific (Fig 2.29) and represents convergence of air between the almost permanent anticyclone near Easter Island in the eastern Pacific and the more mobile anticyclones to the east of Australia and New Zealand. Detailed studies (Fig 2.29) show that there is a high frequency of cyclogenesis along this cloud band at least as far south as 40°S with a maximum near 150°W between about 25°S and 35°S. Of course the data are poor and details are uncertain. Some AGCM's reproduce this feature quite well, though it has not yet been much studied even in the models. An exception is a recent paper by James and Anderson (1984) (21) who discuss a similar but less permanent feature that may exist in the S Atlantic and does exist in the ECMWF model. Streten and Zillman (1984) give an extensive description of the climate of the S Pacific Ocean.

Another peculiar feature of the southern oceans, including the S Pacific region, is that the mean latitude of the depression belt around Antarctica varies semi-annually (Fig 2.30). This fluctuation is related to a more general semi-annual variation of circulation in the S Hemisphere (Hsu and Wallace (1976)).

2.6 Summary

This long lecture has tried to show that the climate of the world and its variability is very interesting, complex and nowhere completely understood. It is a fitting subject for the World Climate Research Programme. I recommend the 15 volumes of the "World Survey of Climatology" (which include Schwerdtfeger (1976)) for those who require factual detail on specific regions.

Acknowledgement

I have drawn freely from Mr D Jones 1982 Advanced Lecture 2.

LECTURE 2 - REFERENCES

BERKOFSKY, L AND BERTONI, E A (1955) Mean topographic charts for the entire earth. Bull Amer Met Soc, 36, pp 350-354.

CHIU, L S (1983) Variation of Antarctic Sea Ice: an update. Mon Weath Rev, 111, pp 578-580.

CRITCHFIELD, H J (1983) General Climatology. 4th Edition. Prentice-Hall.

DARWIN, C (1979) reprint of "Voyage of the Beagle" (p. 235). J M Dent.

FOLLAND, C K (1983) Regional-scale interannual variability of climate - a north west European perspective. Met Mag, 112, pp 163-183.

GADGIL S, JOSEPH, P V, and JOSHI, N V (1984) Ocean-atmosphere coupling over monsoon regions. Nature, 312, pp 141-143.

HAHN, D G (1981) Summary requirements of GCM's for observed snow and ice cover data. In Snow Watch 1980 Ed: G Kukla et al. World Data Centre A for Glaciology.

HASTENRATH, S, WU, M-C, and CHU, P-S (1984) Toward the monitoring and prediction of N E Brazil droughts. Q J Roy Met Soc, 110, pp 411-425.

HOFLICH, O (1984) Climate of the S Atlantic Ocean. In World Survey of Climatology; Climate of the Oceans. Ed: H VAN LOON, Vol 15, pp 1-192.

HSU, C P F and WALLACE, J M (1976) The global distribution of the annual and semiannual cycles in sea level pressure. Mon Weath Rev, 104, pp 1597-1601.

JACOBS, I (1958) Met Abh, Inst Met Geoph, Freien Univ, Berlin, 4, No. 2, 47 pp (Part I); 121 pp (Part II).

JAMES, I N AND ANDERSON, D L T (1984) The seasonal mean flow distribution of large-scale weather systems in the southern hemisphere: the effects of moisture transports. Q J Roy Met Soc, 110, pp 943-966.

KOPPEN, W (1931) Grundriss der Klimakunde, Berlin.

KORNFELD, J AND HASLER, A F (1969) A photographic summary of the earth's cloud cover for the year 1967. J Appl Met, 8, pp 687-699.

LAMB, H H (1972) Climate Present, Past and Future. Volume 1. Methuen.

LEJENAS, H AND OKLAND, H (1983) Characteristics of northern hemisphere blocking as determined from a long time-series of observational data. Tellus, 35A, pp 350-362.

LETTAU, H H AND LETTAU, K (Eds) (1978) Exploring the World's driest climate. Inst for Env Studies Rep 101, Univ of Wisconsin, pp 182-248.

MANABE, S AND HOLLOWAY, J L (1975) The seasonal variation of the hydrological cycle as measured by a global model of the atmosphere. J Geophys Res, 80, pp 1617-1649.

McALPINE, J R, KEIG, G, and FALLS, R (1983) Climate of Papua New Guinea. CSIRO. Aust Nat Univ Press.

MEIER, M F (1983) Snow and ice in a changing hydrological world. Hydrol. Sci. Journ. 28, pp.3-22.

MINHINICK, J H AND FOLLAND, C K (1984) The Meteorological Office Historical Sea Surface Temperature Dataset (MOHSST). Met O 13 Branch Memo No 137.

NEWELL, R E AND KIDSON, J W (1984) African mean wind changes between Sahelian Wet and Dry periods. J Clim, 4, pp 27-33.

NEWELL, R E, KIDSON, J W, VINCENT, D G AND BOER, G J (1974) The General Circulation of the Tropical Atmosphere. Vol 2, pp 151, MIT Press.

NEWELL, R E, VINCENT, D G, DOPPLICK, T G, FERRUZA, D AND KIDSON, J W (1969) The energy balance of the global atmosphere. In "The Global Circulation of the Atmosphere", Roy Met Soc, pp 42-90.

PALMEN, E AND NEWTON, C W (1969) Atmospheric circulation systems. Their structure and physical interpretation. International Geophysics series, Vol 13, Academic Press, New York and London.

PERRY, A H AND WALKER, J M (1977) The ocean-atmosphere system. Longman, London and New York.

RIDDAWAY, R W (1980) Monsoons. In Advanced Lectures, 1980, on "Some aspects of dynamical meteorology". Unpublished: available from Met Office College.

RIEHL, H (1979) Climate and weather in the tropics. Academic Press.

SCHWERDTFEGER, W (1976) Introduction to the Climate of Central and South America. In World Survey of Climatology; Climate of C and S America. Ed: W SCHWERDTFEGER, Vol 12.

STRETEN, N A (1973) Some characteristics of satellite-observed bands of persistent cloudiness over the southern hemisphere. Mon. Weath. Rev., 101, pp.486-495.

STRETEN, N A (1980) Some synoptic indices of the southern hemisphere mean sea level circulation 1972-1977. Mon Weath Rev, 108, pp 18-36.

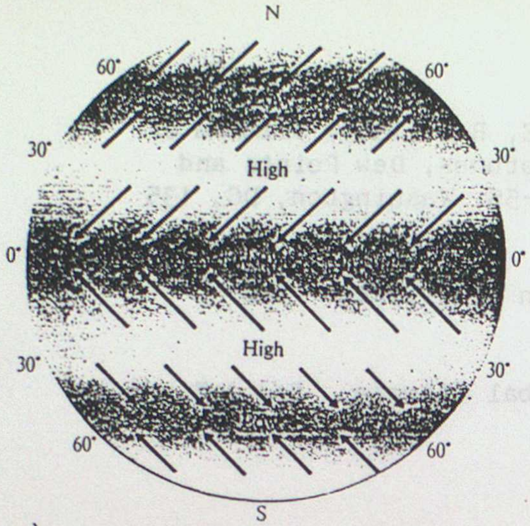
STRETEN, N A AND TROUP, A J (1973) A synoptic climatology of satellite observed cloud vortices over the southern hemisphere. QJR Met Soc, 99, pp 56-72.

STRETEN, N A AND ZILLMAN, J W (1984) Climate of the South Pacific Ocean. In World Survey of Climatology; Climate of the Oceans. Ed: H VAN LOON, Vol 15, pp 263-429.

TALJAARD, J J, VAN LOON, H, CRUTCHER, H L and JENNE, R L (1969) Climate of the Upper Air: Southern Hemisphere, Vol I. Temperatures, Dew Points and Heights at Selected Pressure Levels. NAVAER 50-1C-55, Washington, DC, 135 pp.

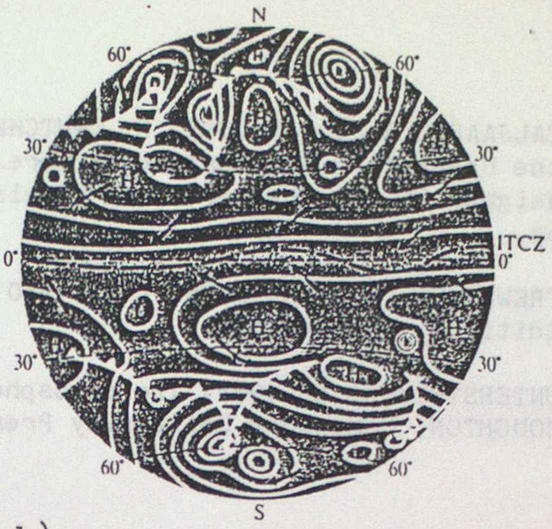
TREWARTHA, G T AND HORN, L H (1980) An introduction to climate. 5th Edition. McGraw-Hill.

UNTERSTEINER, N (1984) The cryosphere. In The Global Climate. Ed: J T HOUGHTON, Cambridge University Press.



a)

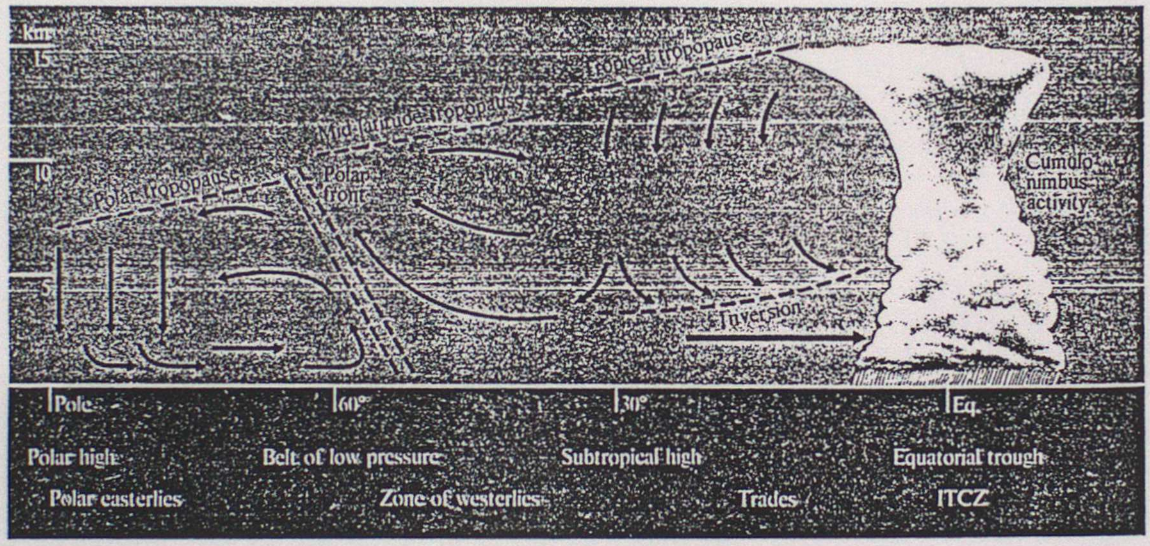
Idealized pattern of pressure and winds over the surface of the globe.



b)

Typical distribution of pressure systems, winds and fronts near an equinox.

Figure 2.1 After Perry and Walker (1977)



Schematic representation of the atmospheric general circulation, in vertical section from polar regions to the equator.

Figure 2.2 After Perry and Walker (1977)

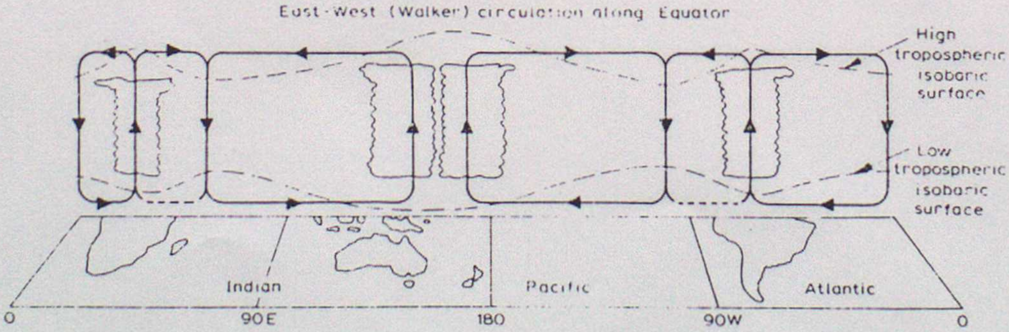


Figure 2.3 From Streten and Zillman (1984)

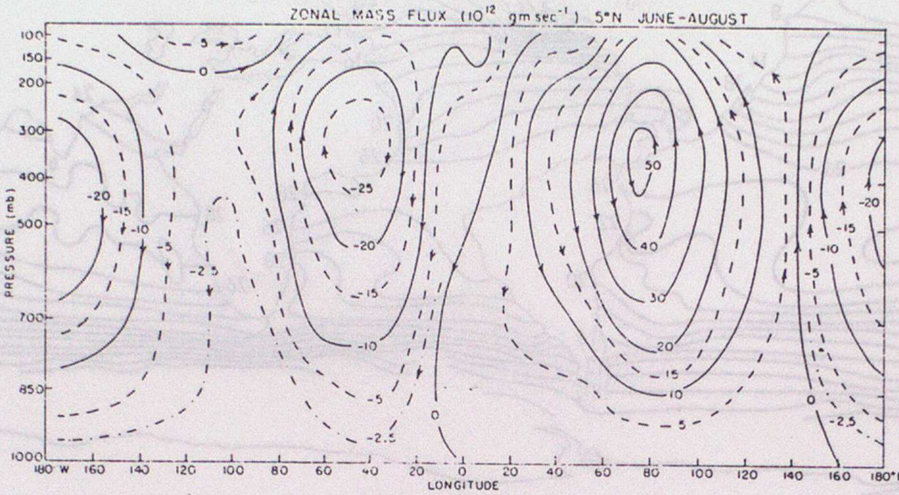
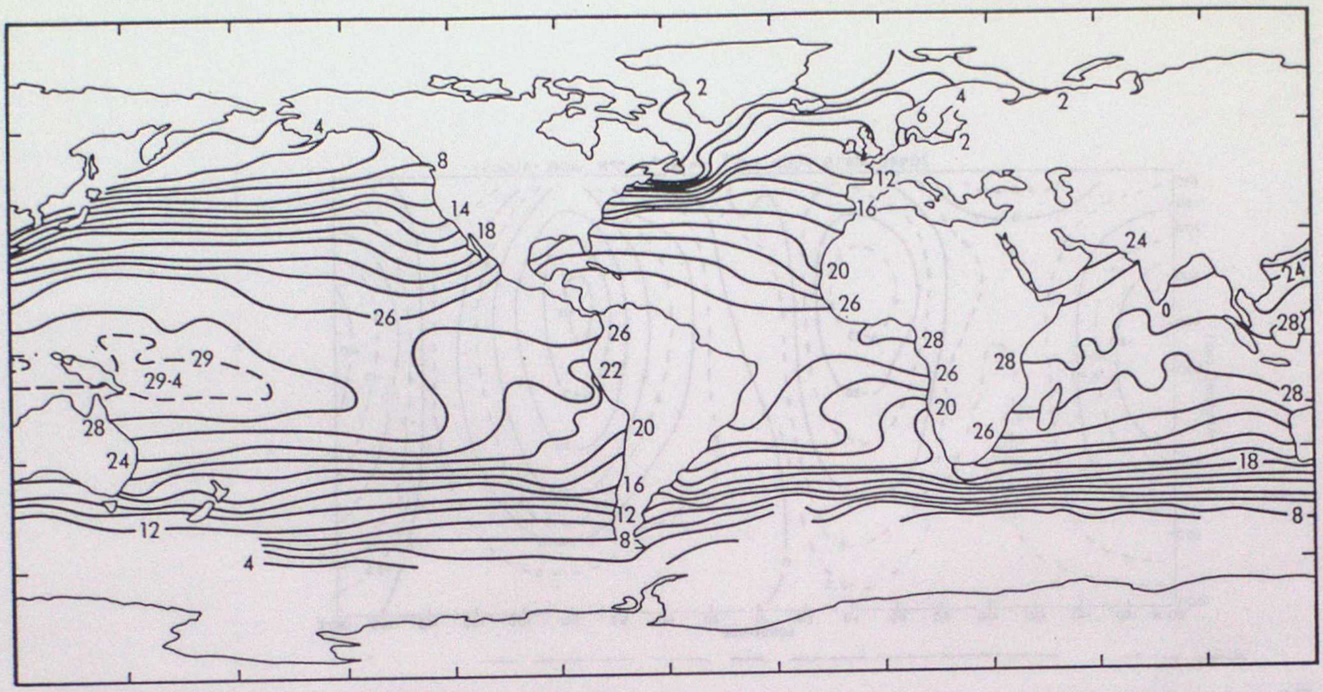
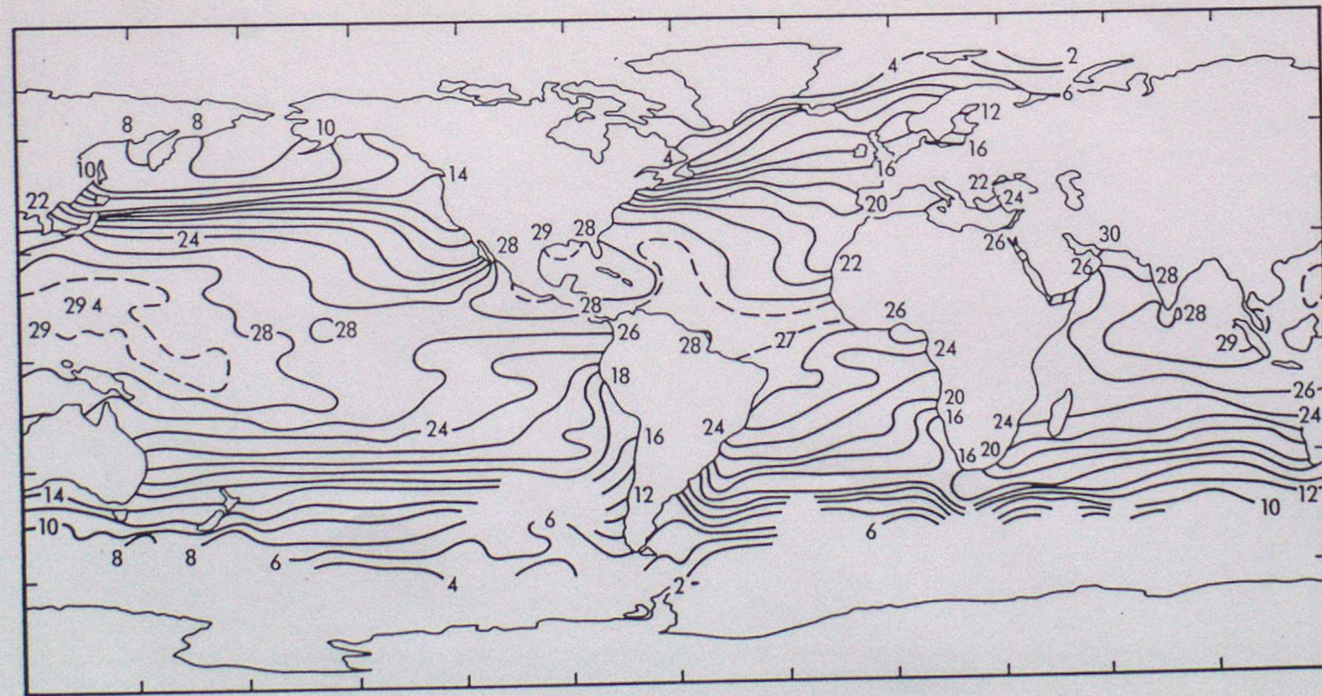


Figure 2.4 A representation of the zonal mass flux averaged over the region 0-10°N for June to August. Contours do not correspond to streamlines, but give a fairly good representation of the velocity field associated with the Walker circulation.
From Newell et al (1974)



Average sea surface temperature (1951-80), January (Based on MOHSST 3)



Average sea surface temperature (1951-80), July (Based on MOHSST 3)

DO 3723

Fig. 2.5

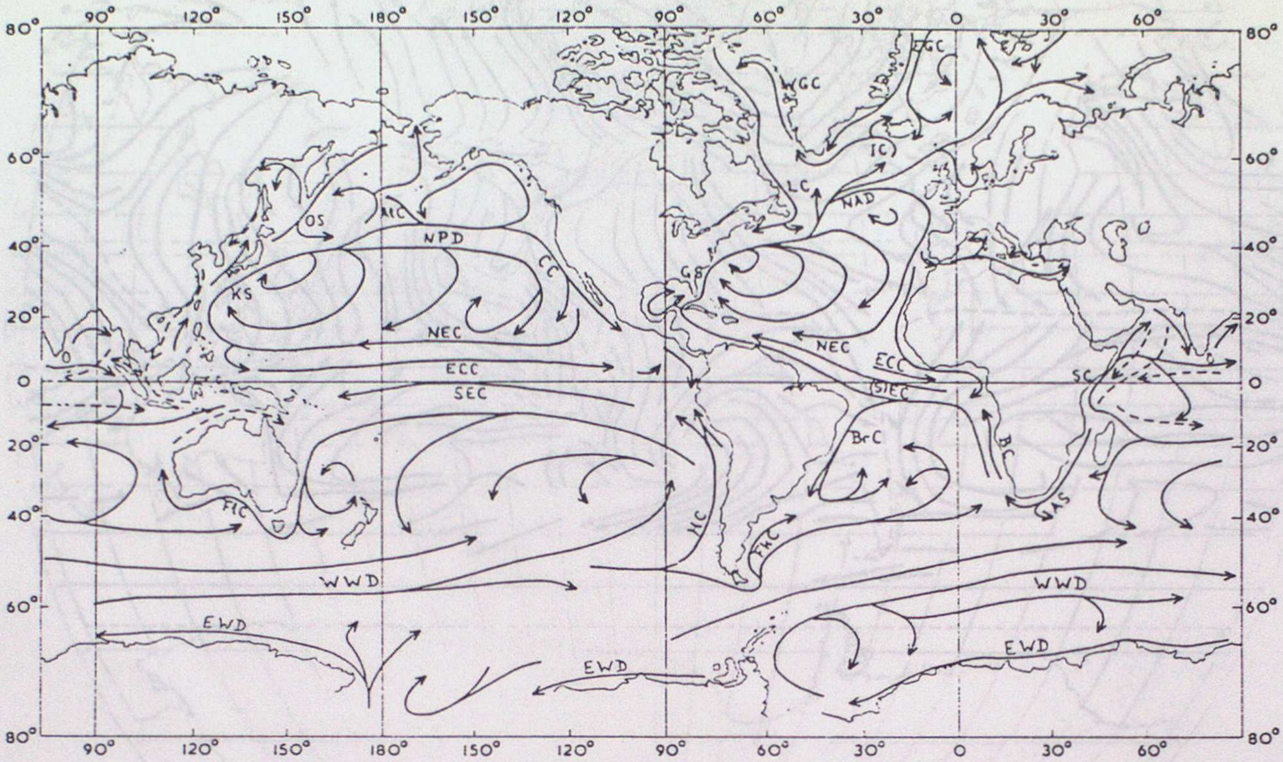


Figure 2.6 Main surface currents of the oceans. From Lamb(1972). The abbreviations for the names of the currents are expanded in table 8.2 of Lamb(1972)

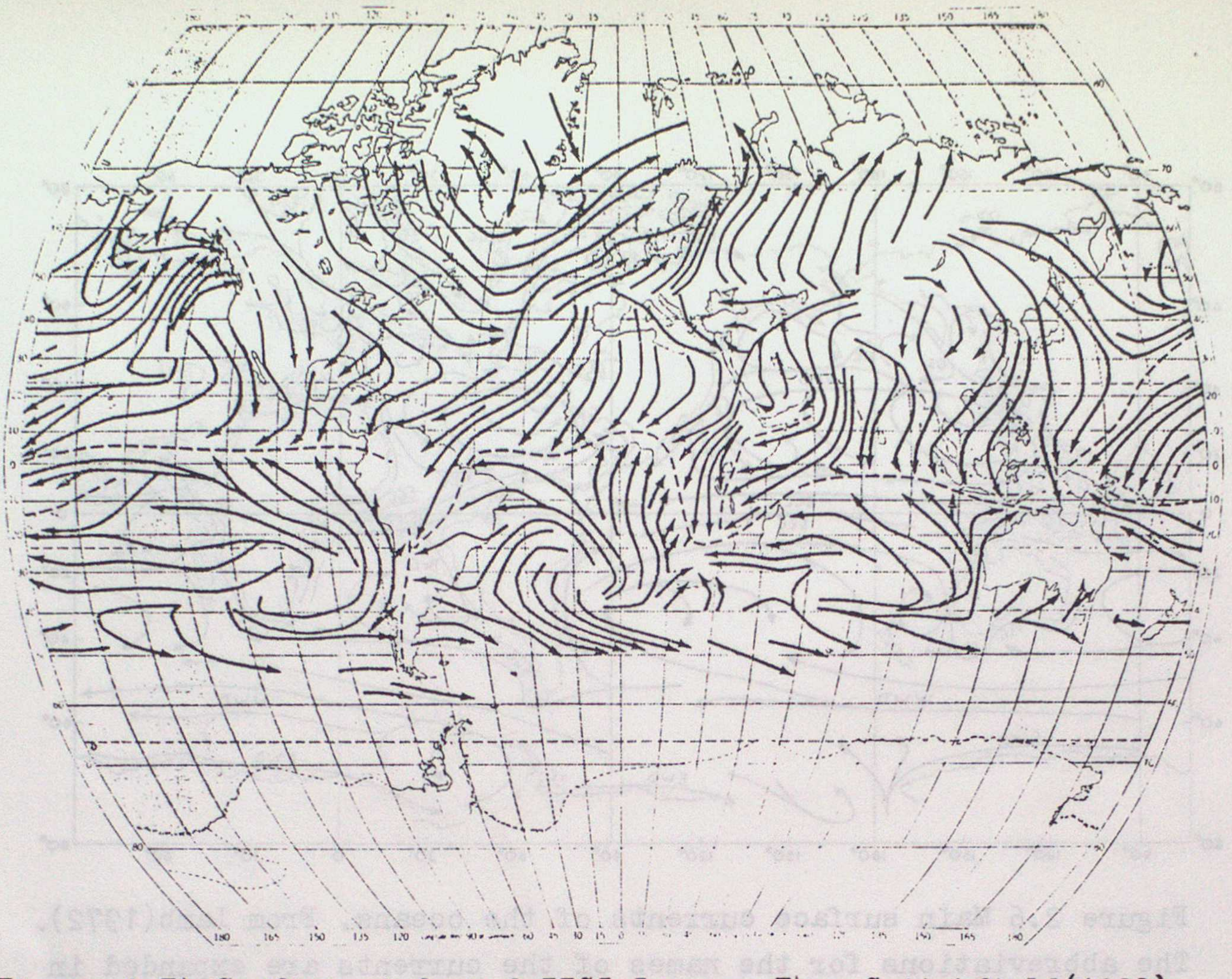


Figure 2.7A Surface winds and ITCZ January.From Critchfield(1983)

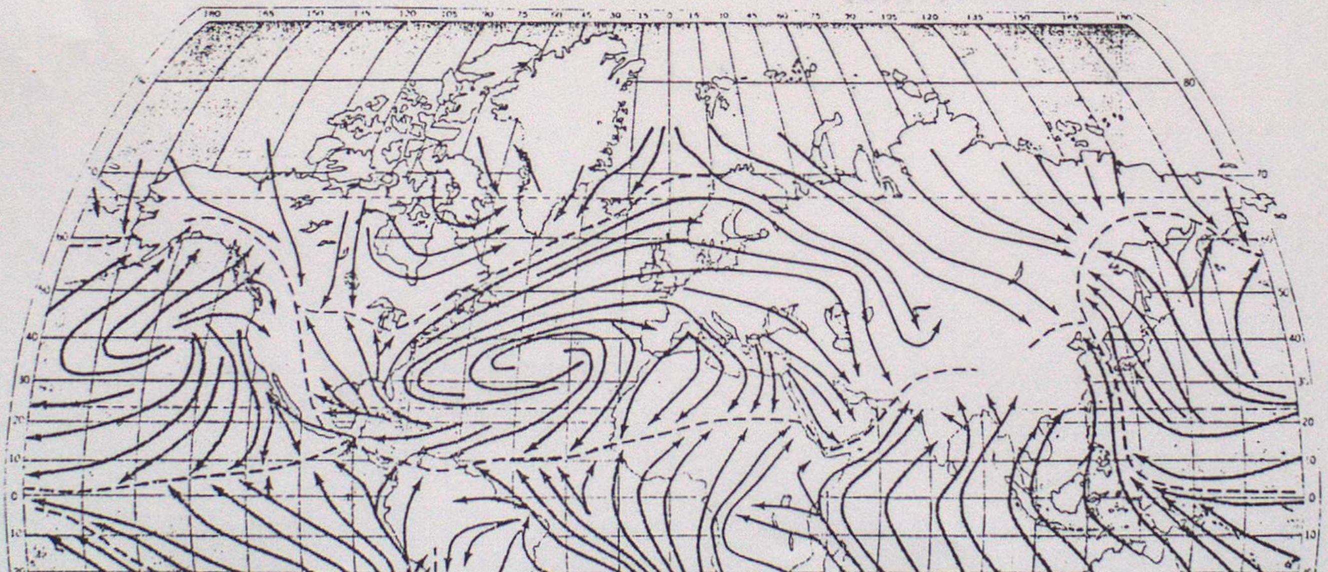


Figure 2.7B Surface winds and ITCZ July.From Critchfield(1983)

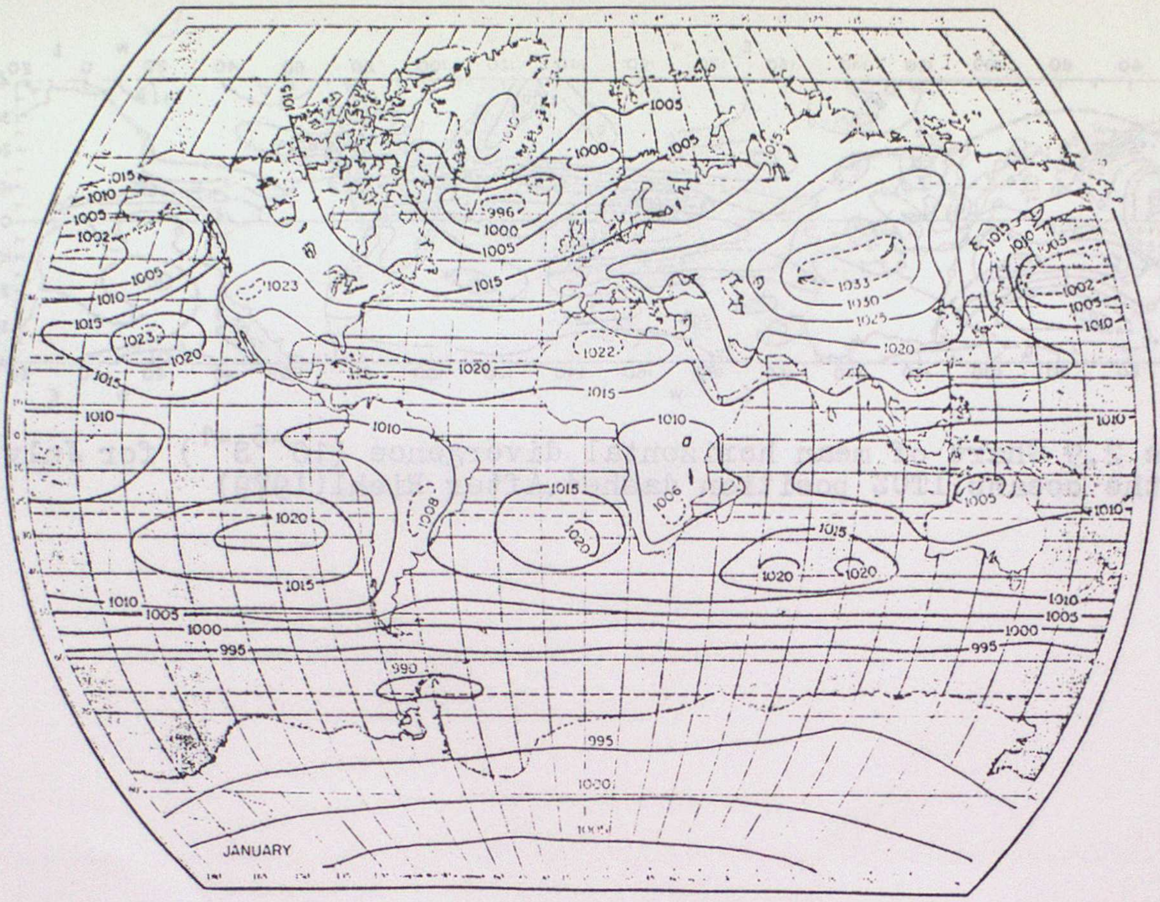


Figure 2.8A World sea-level pressure (mbs) in January. (Modified Van der Grinten projection.)
From Critchfield(1983)

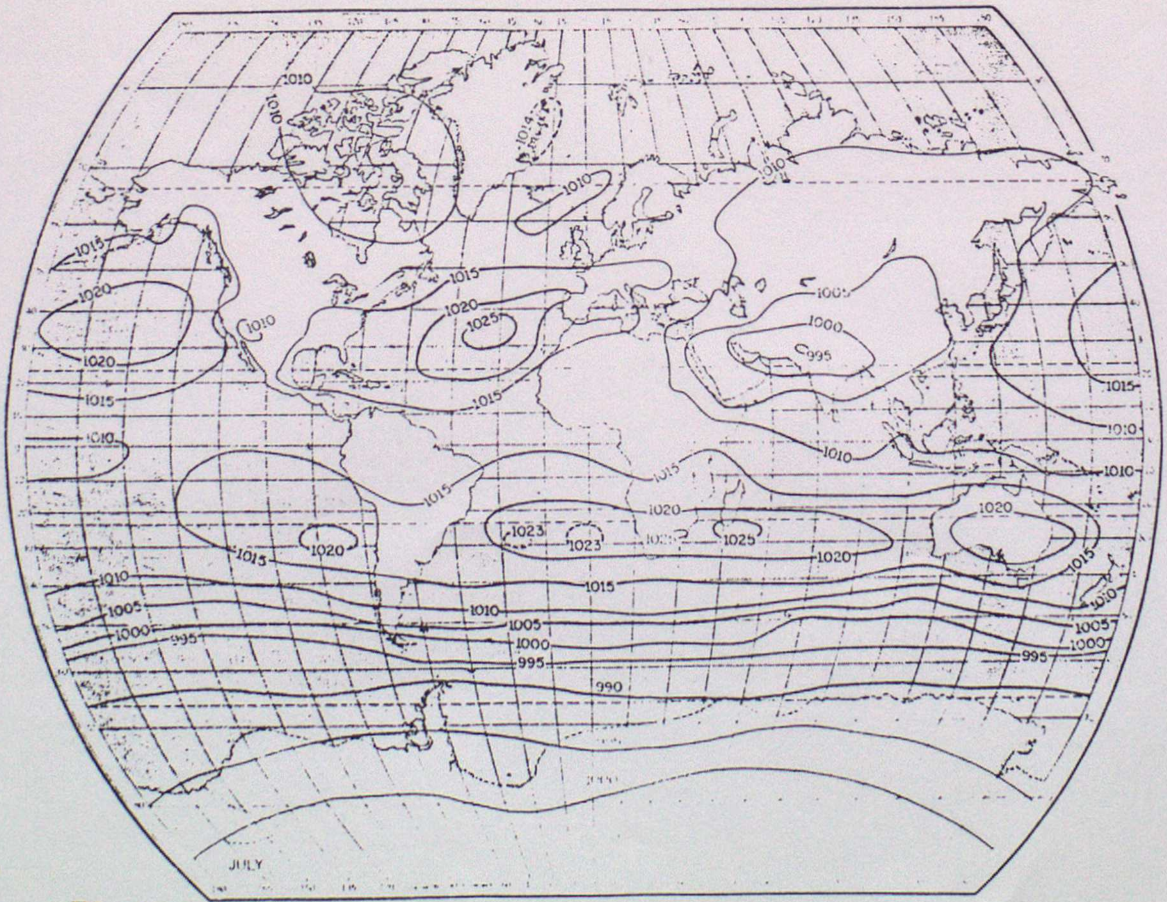


Figure 2.8B World sea-level pressure (mbs) in July. (Modified Van der Grinten projection.)
From Critchfield(1983)

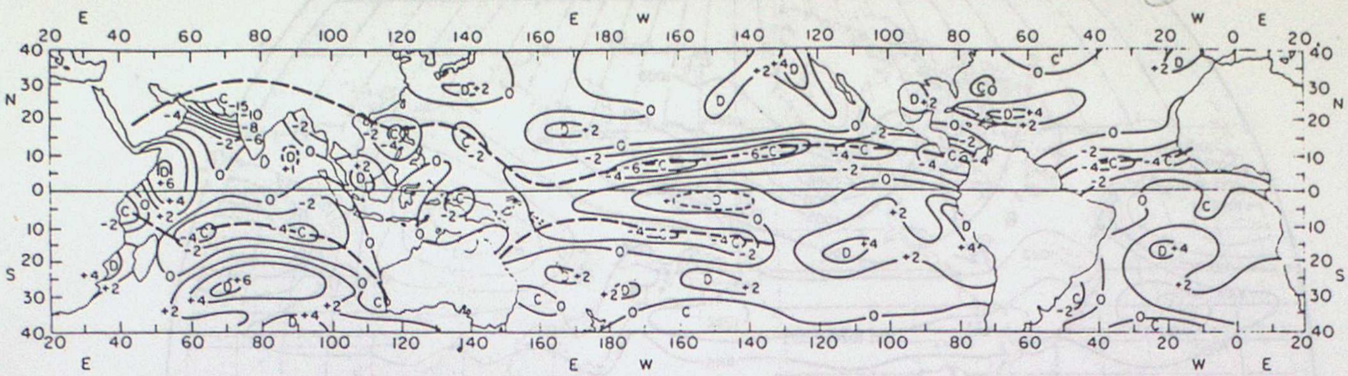


Figure 2.9 Chart of mean horizontal divergence ($10^{-6}S^{-1}$) for July over the oceans.ITCZ position dashed.After Riehl(1979)

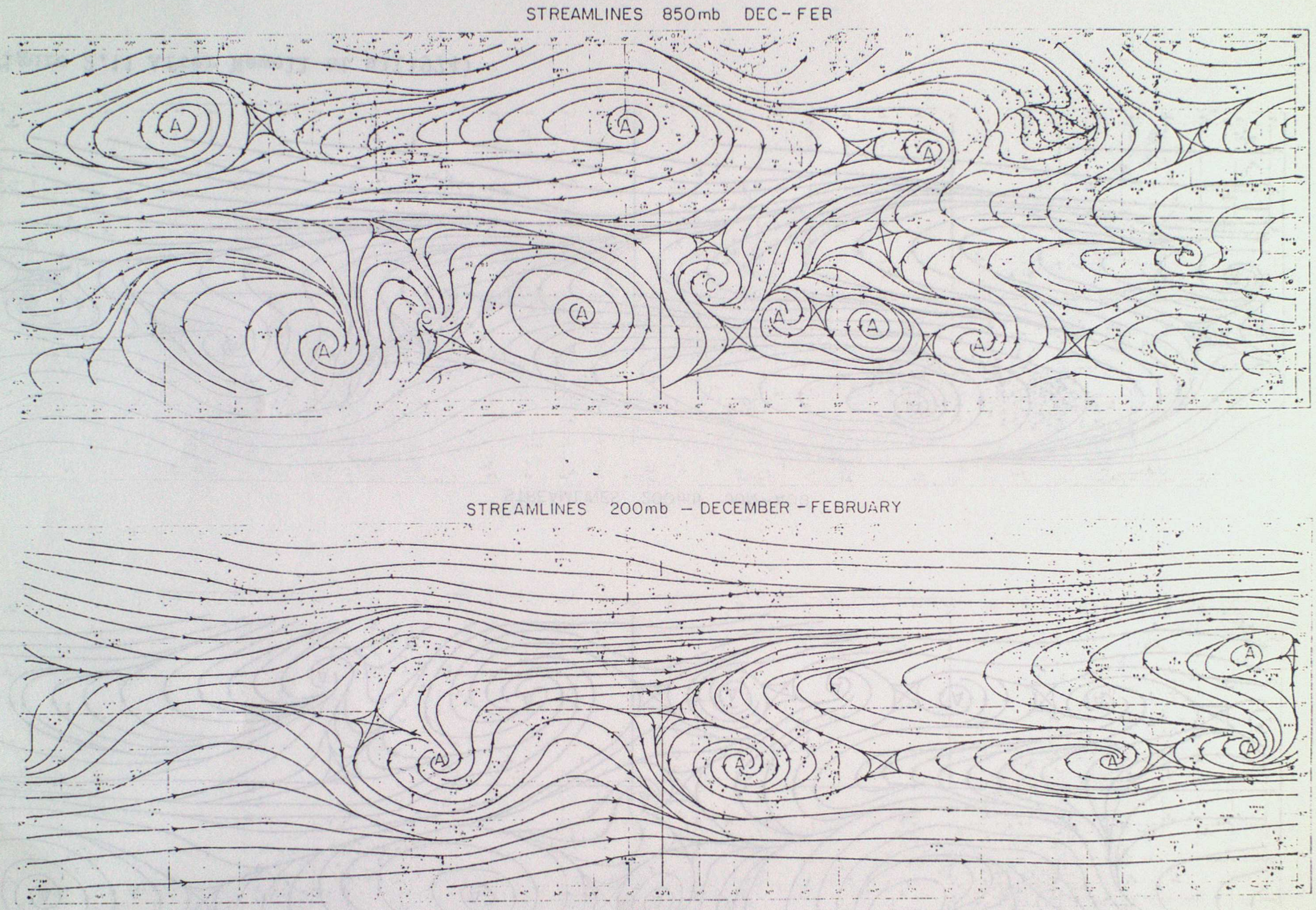


Figure 2.10 After Newell et al (1974)

STREAMLINES 850mb JUN-AUG

STREAMLINES 200mb JUN-AUG

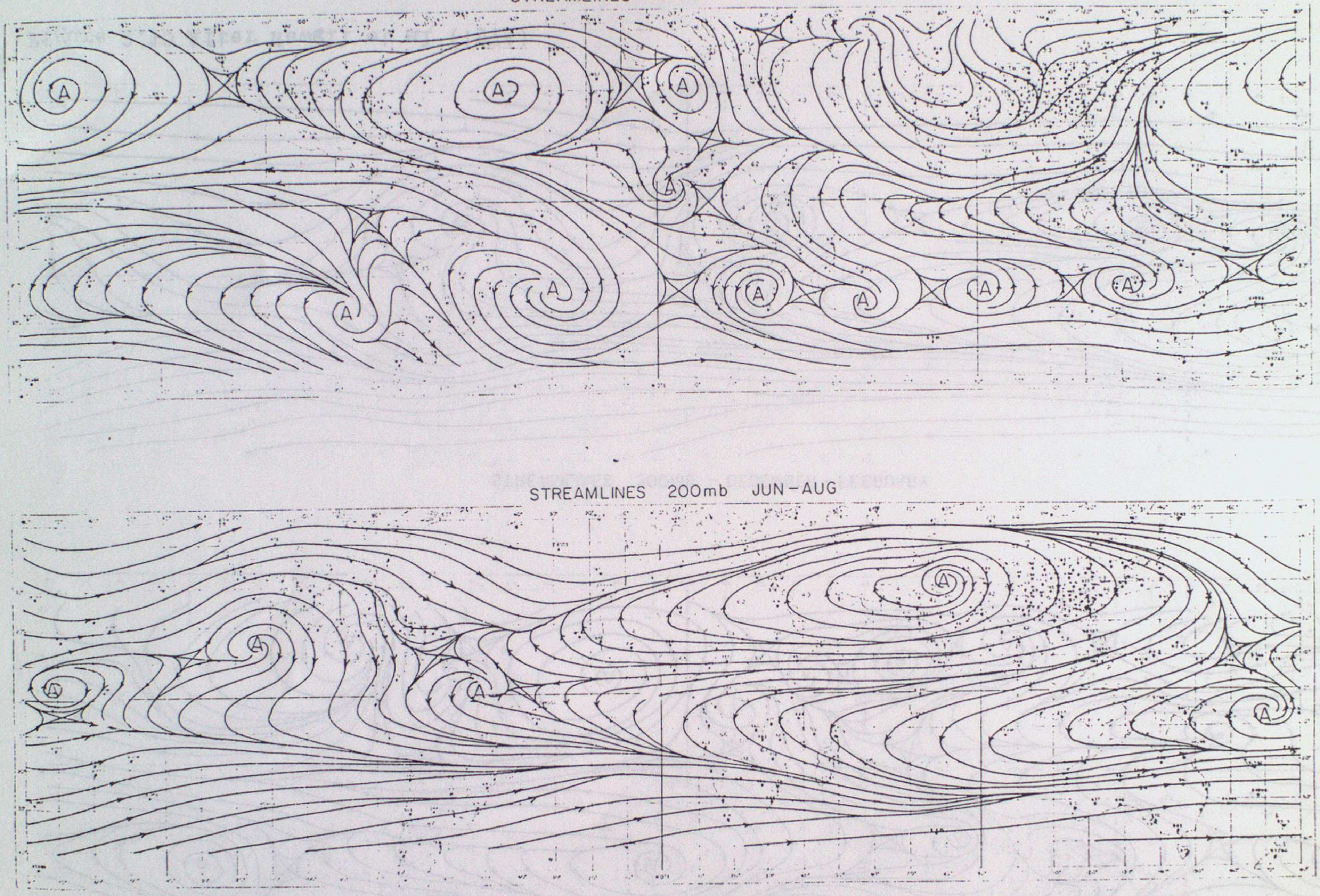
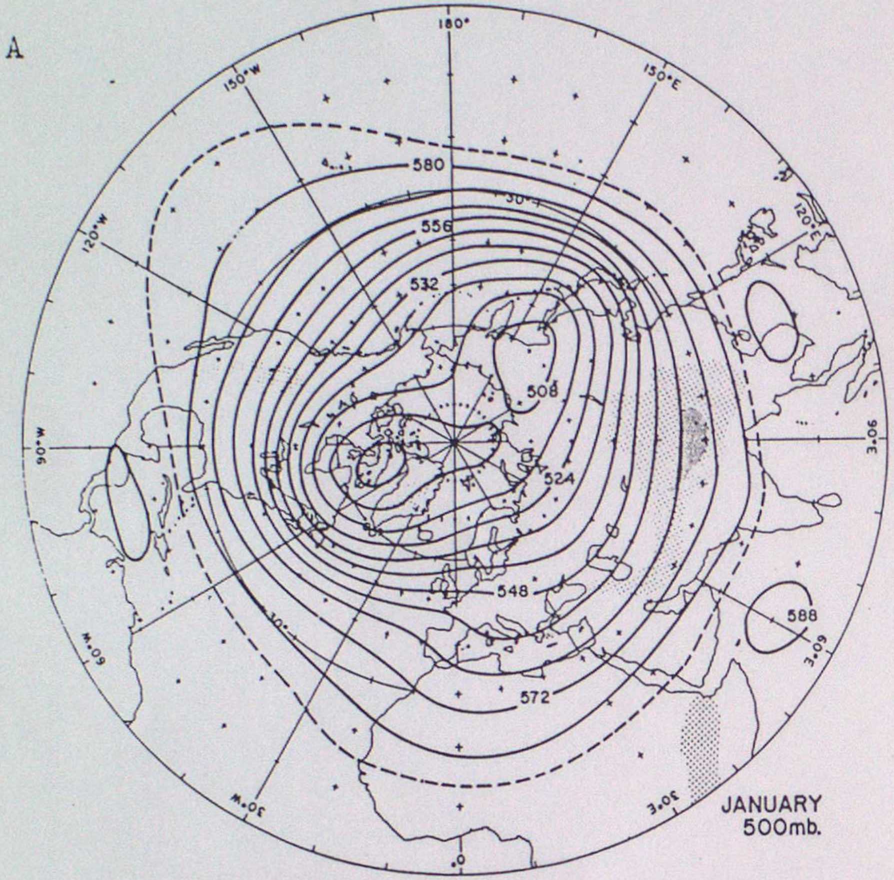


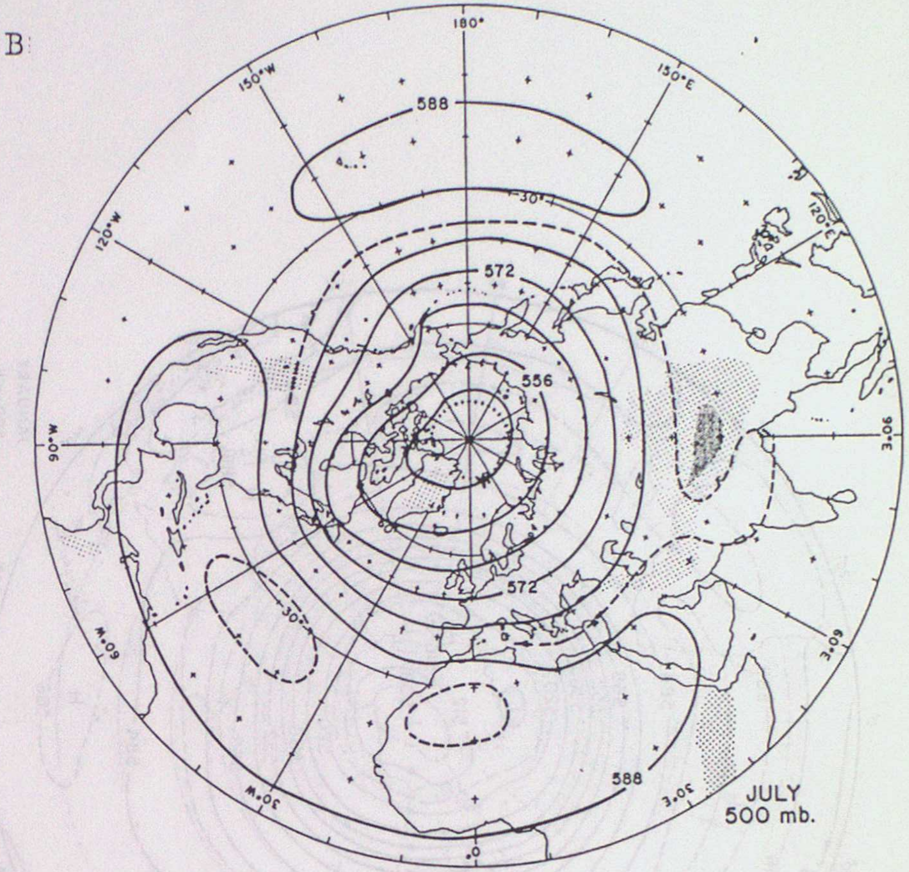
Figure 2.11 After Newell et al(1974)

A



Mean 500-mb contours in January (winter), Northern Hemisphere. Redrawn at 80-m intervals from I. Jacobs (1958). Light and heavier stippling show regions where elevations are above 1.5 km and 5 km (smoothed over 5° latitude-longitude tessera), from Berkofsky and Bertoni (1955).

B



Mean 500-mb contours in July (summer), Northern Hemisphere. (Redrawn, from I. Jacobs, 1958.)

Figure 2.12

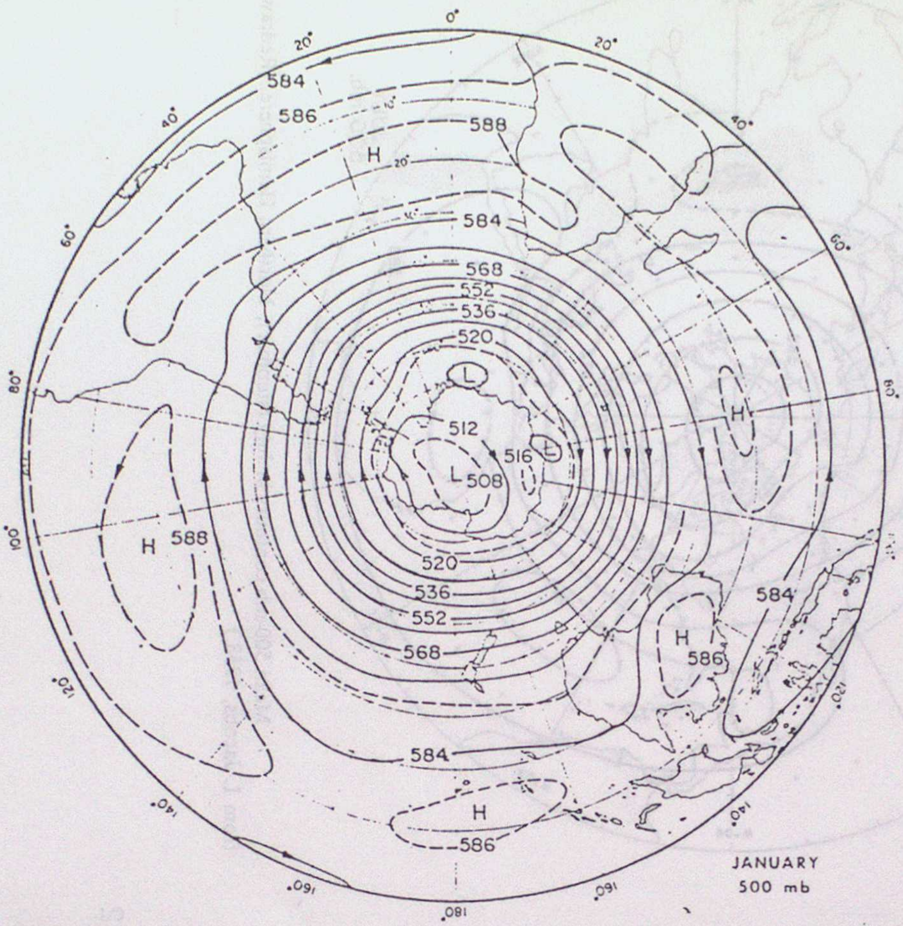


Figure 2.13 . Mean 500-mb contours (80-m interval) in January (summer), Southern Hemisphere. (After Taljaard *et al.*, (1969))

(*a*) Snow and ice in the northern hemisphere at their minimum and maximum extent. (*b*) Snow and ice in the southern hemisphere at their minimum and maximum extent (from Meier, 1983). It should be noted that the summer sea ice extent in Fig. 8.1 appears larger than the corresponding value shown in Table 8.1. The reason is that sea ice 'edge' or 'boundary' are definable only in the context of the method by which the data were obtained or by the purpose for which the data are to be used.

The best set of data for Antarctic sea ice was derived from passive microwave images (ESMR on NIMBUS 5, beginning in December 1972). For reasons of accuracy and unambiguous interpretation of these data, the ice limit was taken to be at concentrations of > 15%, which tends to make the total ice extent appear large. Ships reporting the encounter with ice are likely to place the 'ice margin' at a far higher concentration.

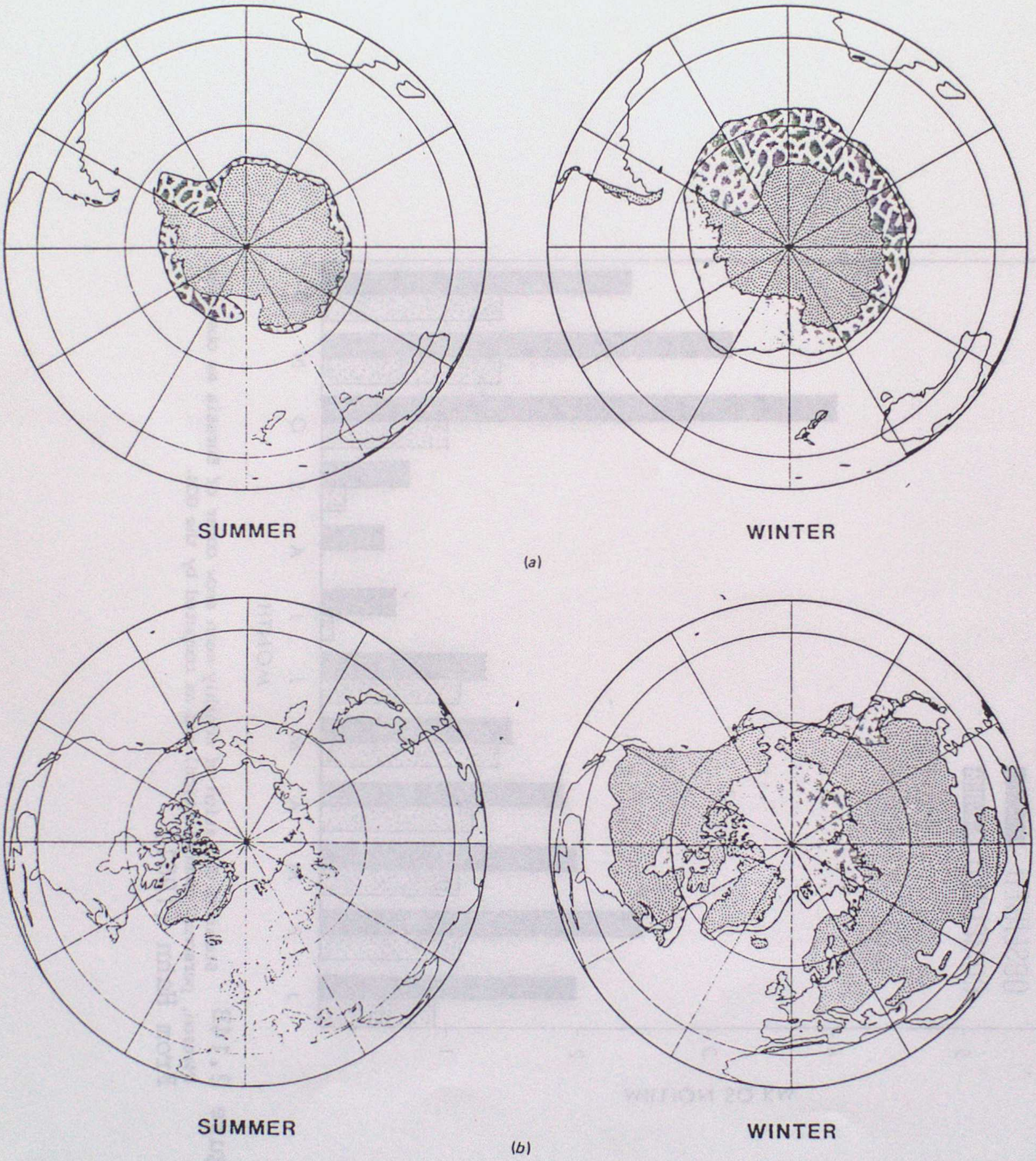


Figure 2.14A After Untersteiner (1984)

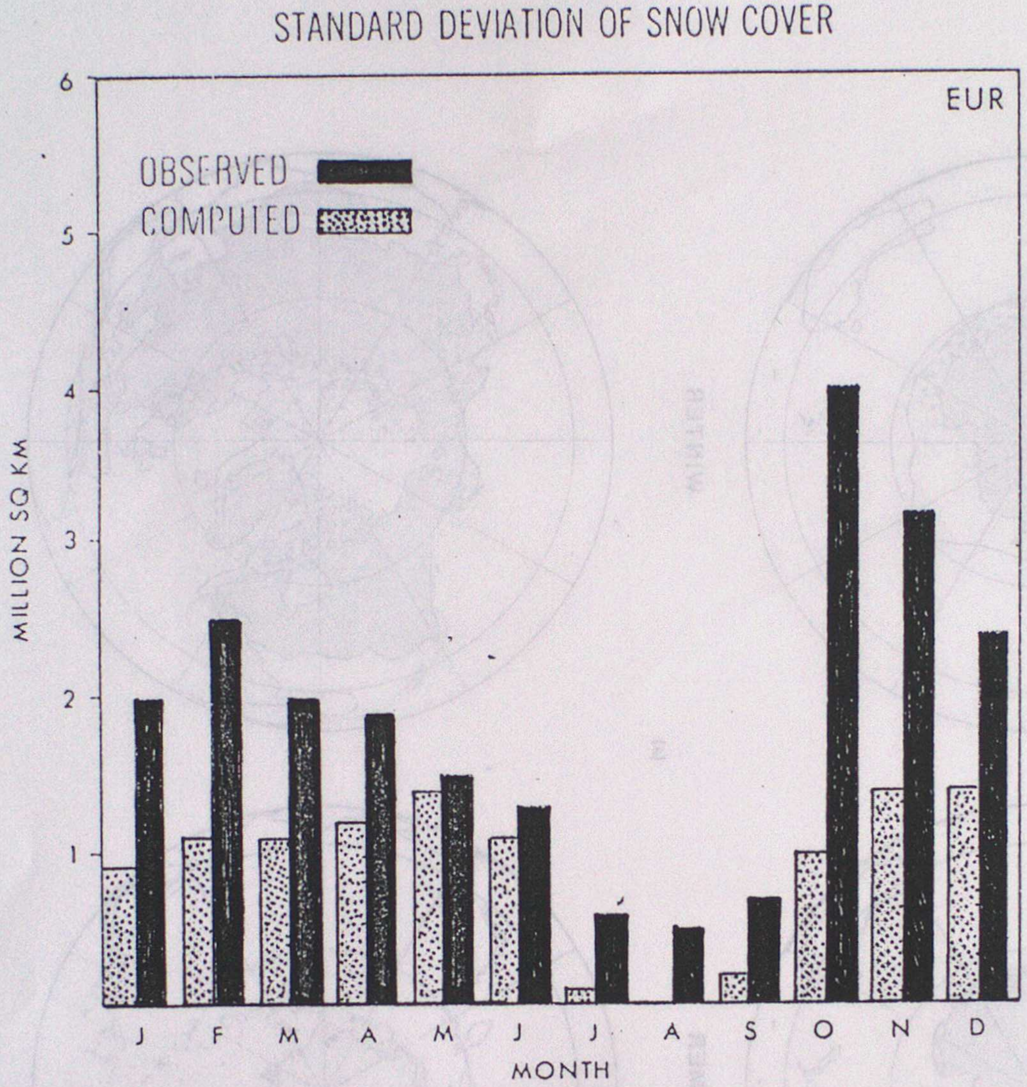


Figure 2.14B: Standard deviation of monthly mean snow cover of Eurasia as observed (Matson, personal communication) and as computed by the GCM. From Hahn (1981)

Figure 2.15 From Critchfield (1983)

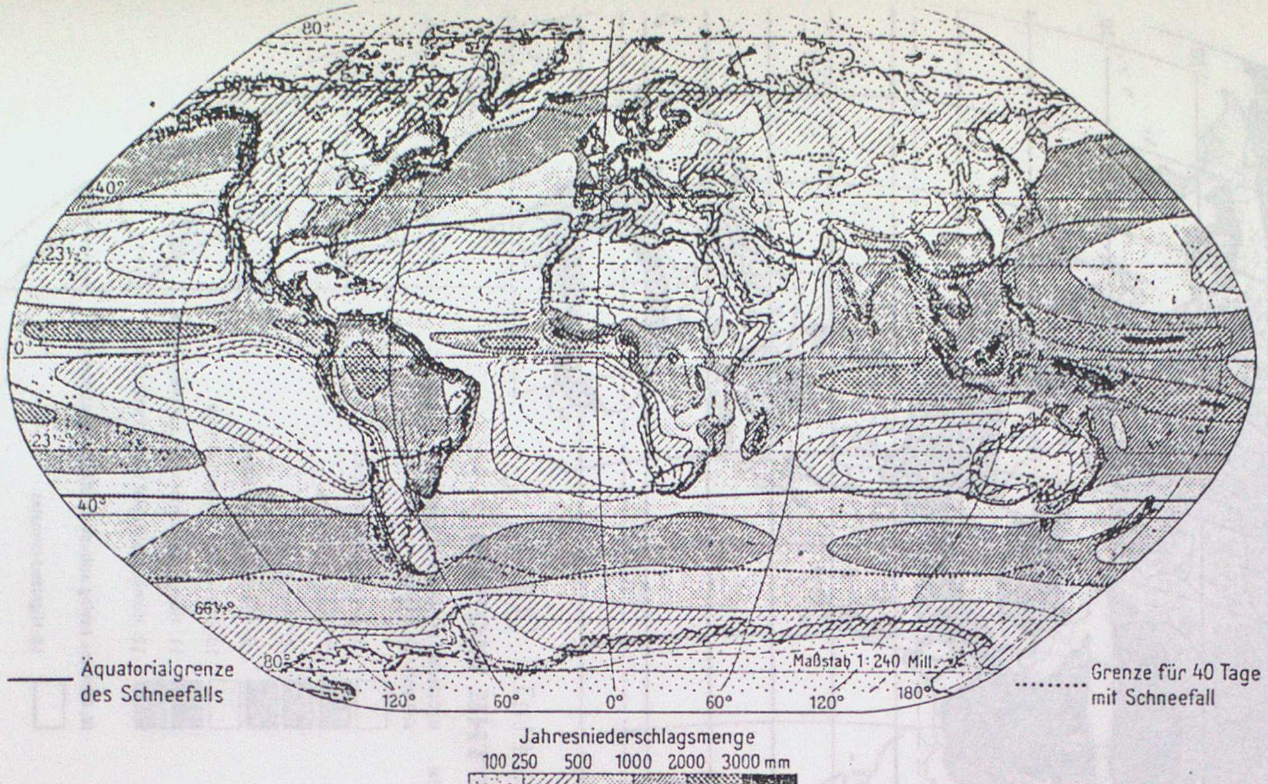


Figure 2.16 Average annual precipitation over the earth (Jahresniederschlagsmenge) in mm. Also shown are the equatorward limits of snowfall (Äquatorialgrenze des Schneefalls) and the limits for 40 days with snowfall (Grenze für 40 Tage mit Schneefall). From Trewartha and Horn (1980)

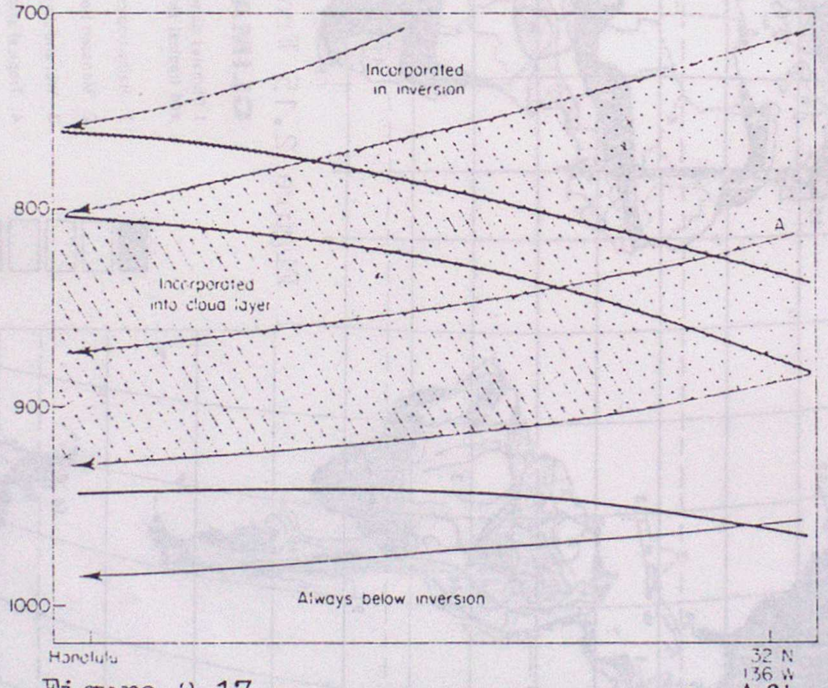


Figure 2.17. The flow of mass through the trade inversion. After Riehl (1979)

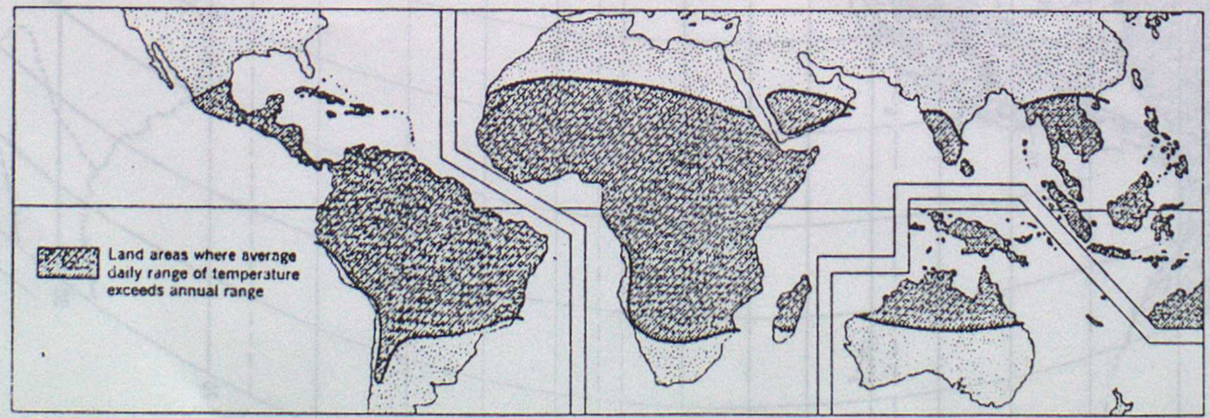


Figure 2.13 A distinctive feature of most tropical climates is that average *daily* range of temperature exceeds average *annual* range. From Trewartha and Horn (1980)

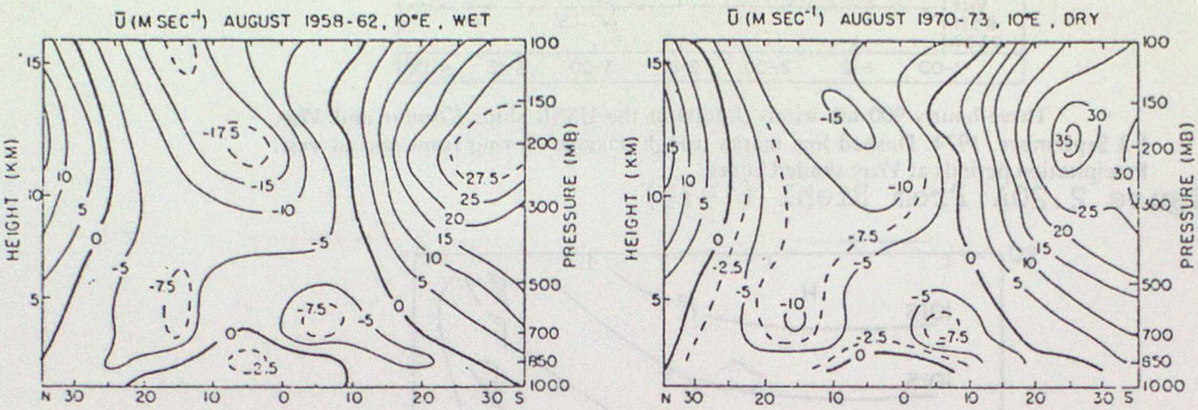


Figure 2.19A. Meridional cross-sections along 10°E of mean zonal wind for August 1958–1962 and August 1970–1973. Units: m s^{-1}
After Newell and Kidson (1984)

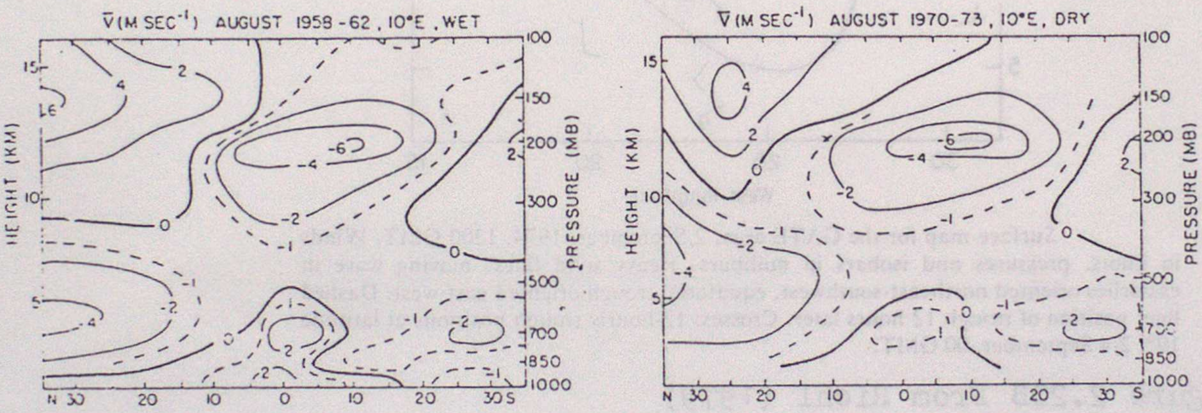
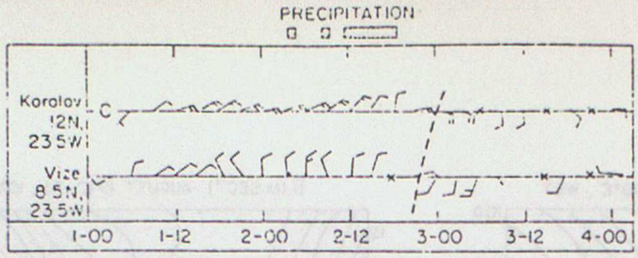


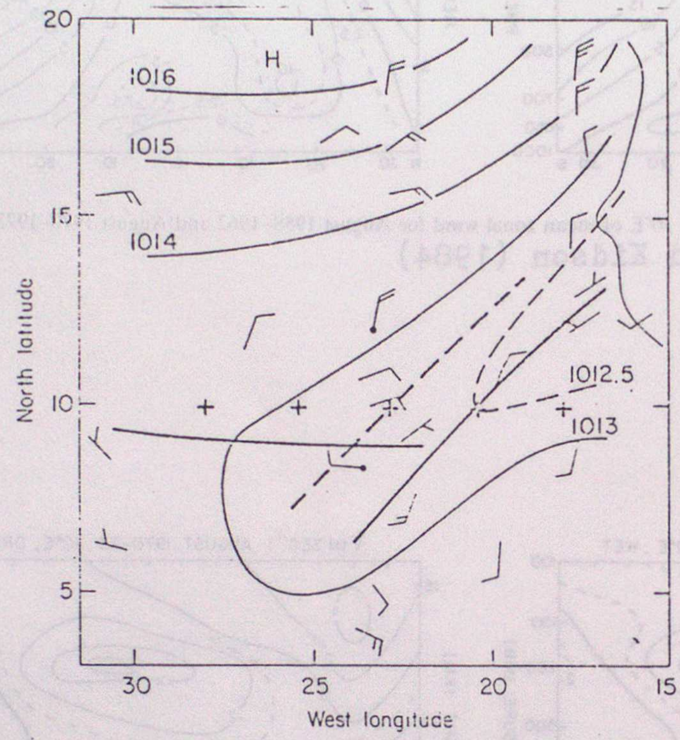
Figure 2.19B Meridional cross sections along 10°E of mean meridional wind for August 1958–1962 and August 1970–1973. Units: m s^{-1}

After Newell and Kidson (1984)



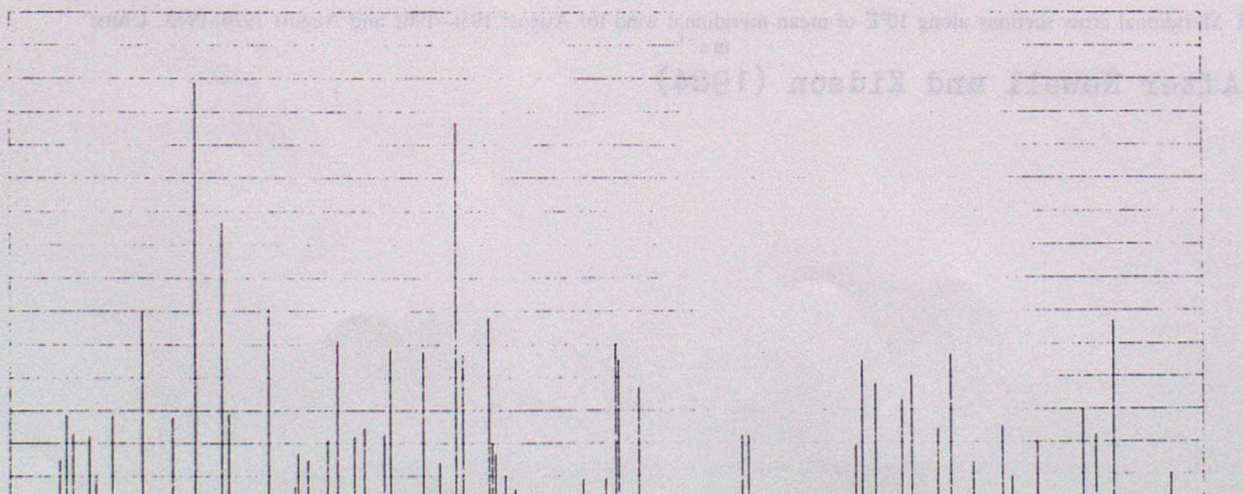
Three-hourly 900 mb winds (knots) at the USSR ships *Korolov* and *Vize*. 1-3 September, 1974. Dashed line marks trough passage moving from east to west. Precipitation periods at *Vize*: shaded boxes.

Figure 2.20A From Riehl (1979)



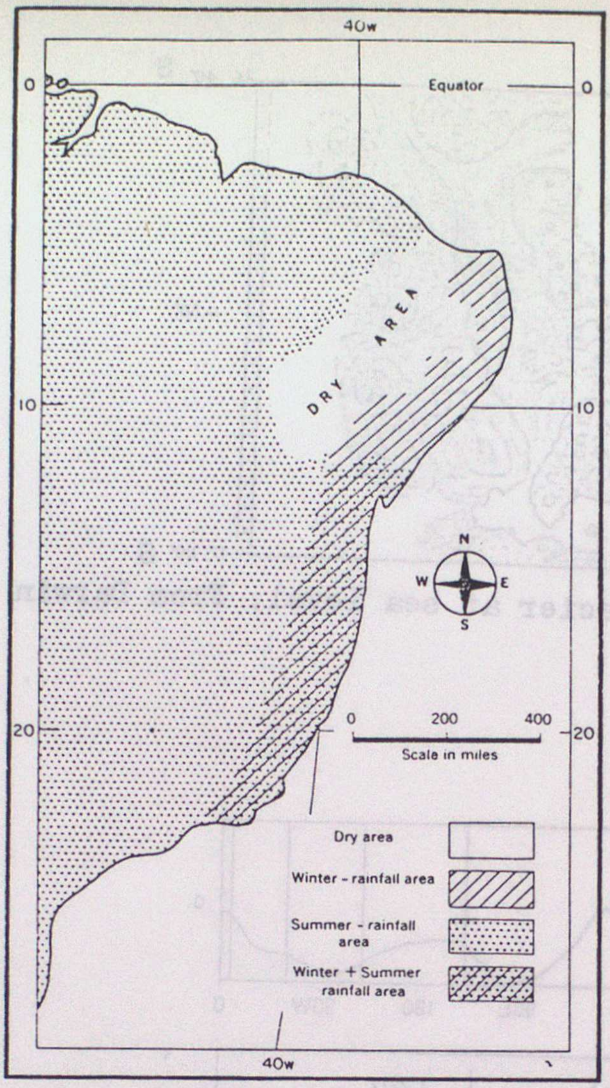
Surface map for the GATE area, 2 September, 1974, 1200 GMT. Winds in knots, pressures and isobars in millibars. Heavy solid lines: moving wave in easterlies oriented northeast-southwest, equatorial trough oriented east-west. Dashed line: position of trough 12 hours later. Crosses: 12-hourly trough positions at latitude 10°, 2-4 September, 00 GMT.

Figure 2.20B From Riehl (1979)



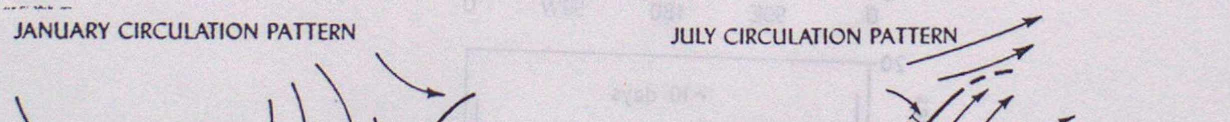
Daily rainfall amounts for one year (1928) at Andagoya, an *Af* station in western Colombia inland from the Pacific coast. Here the average annual rainfall is exceedingly heavy (7089 mm; 282 in), and no month is dry. In 1928, there were only 48 days on which no rain fell. (*After M. Hendl,* Einführung in die physikalische Klimatologie, *Band II*)

Figure 2.21 From Trewartha and Horn (1980)



The dry area in eastern Brazil occupies an intermediate position between the winter-rainfall region to the east and south and the strong summer-rainfall region to the interior and west. It normally escapes the full effects of the disturbances which produce these contrasting seasonal rainfalls.

Figure 2.22 From Trewartha and Horn (1980)



Principal elements of the low-level circulation patterns over eastern and southern Asia in the cold and the warm seasons. (*After Thompson, Watts, Flohn, and others. From Trewartha,* The Earth's Problem Climates, *University of Wisconsin Press, Madison, Wis.*)

Figure 2.23 From Trewartha and Horn (1980)

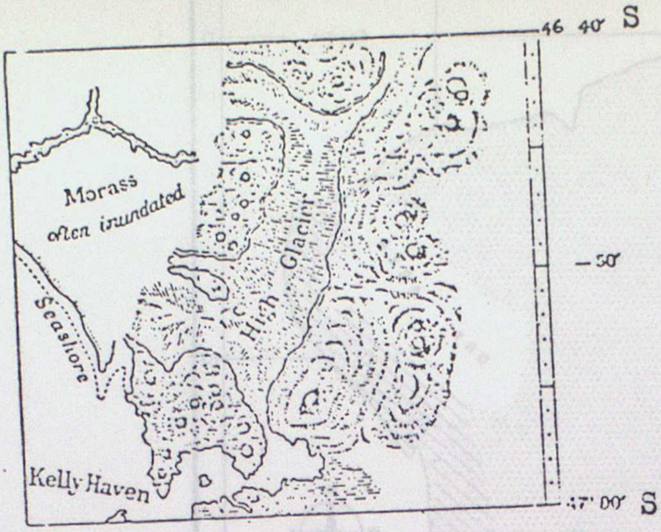
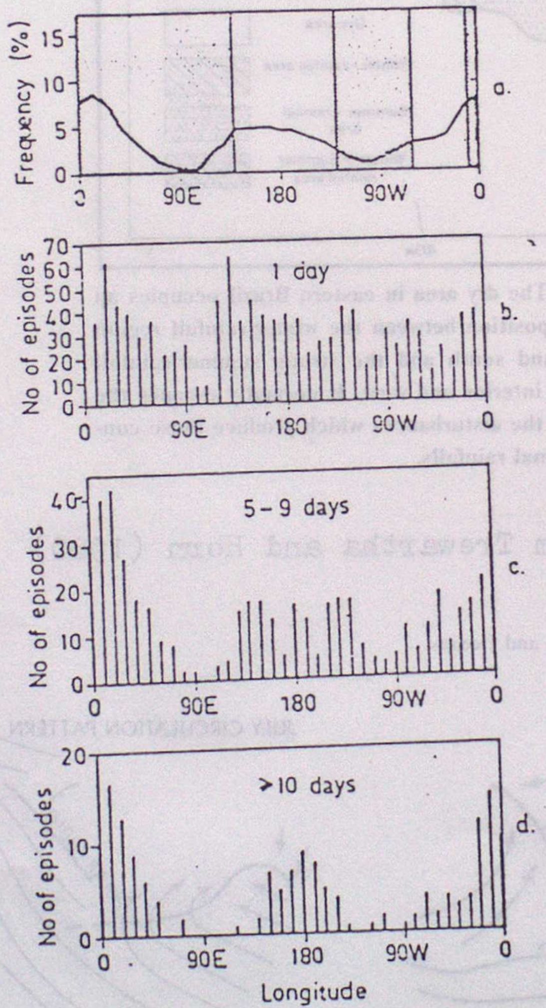
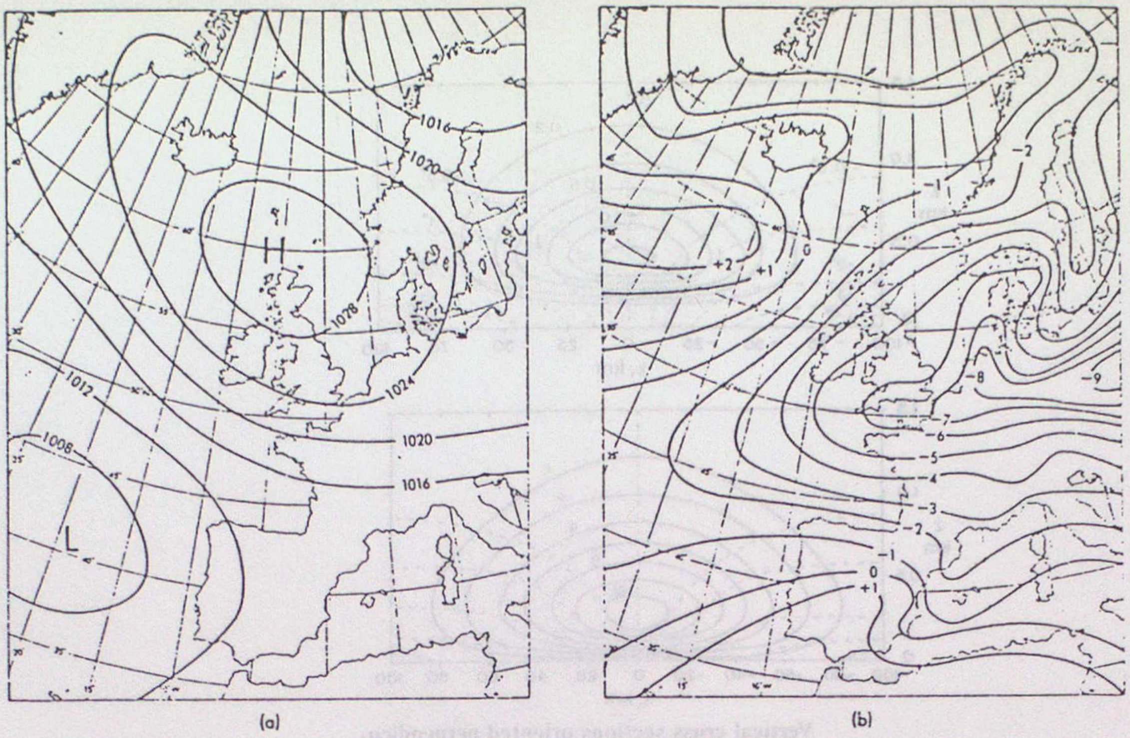


Figure 2.24 Glacier at sea level. From Darwin (1979).

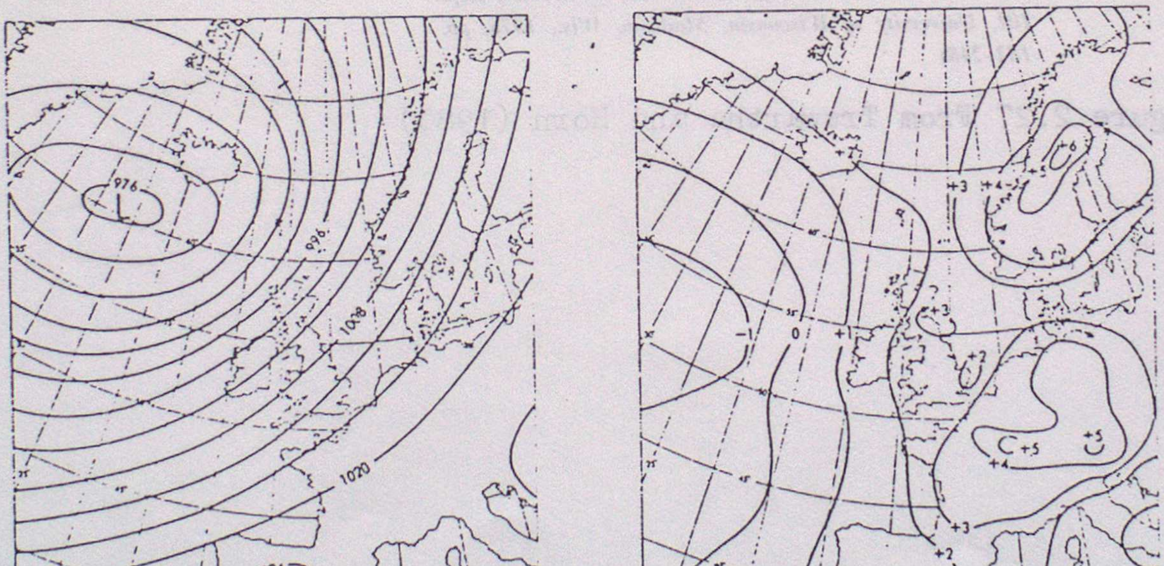


Longitudinal variation of blocked 500-mb flow averaged over the whole year. Continents are shaded. (b–d) Frequency distribution of blocking episodes as a function of longitude, 1950–1979. Duration 1 day (b), 5–9 days (c) and >10 days (d).

Figure 2.25 After Lejenas and Okland (1983)

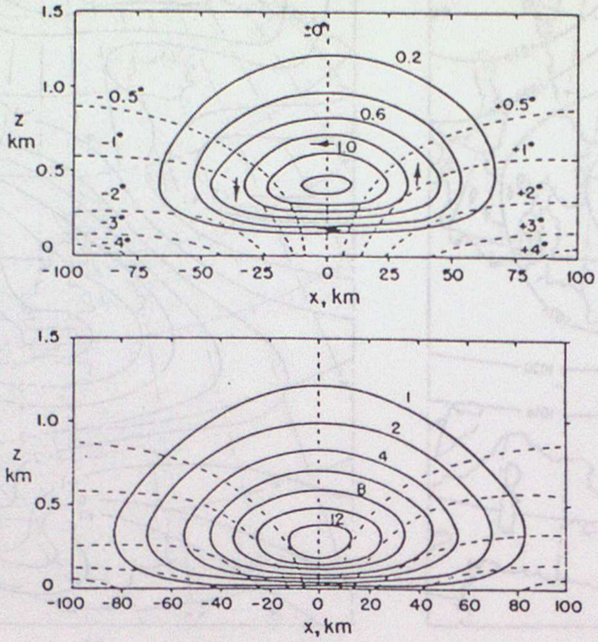


a) Average pressure (mb) at mean sea level, January 1963.
b) Average screen-level temperature anomaly (°C), January 1963. (Anomalies mainly from 1931–60 station averages.)



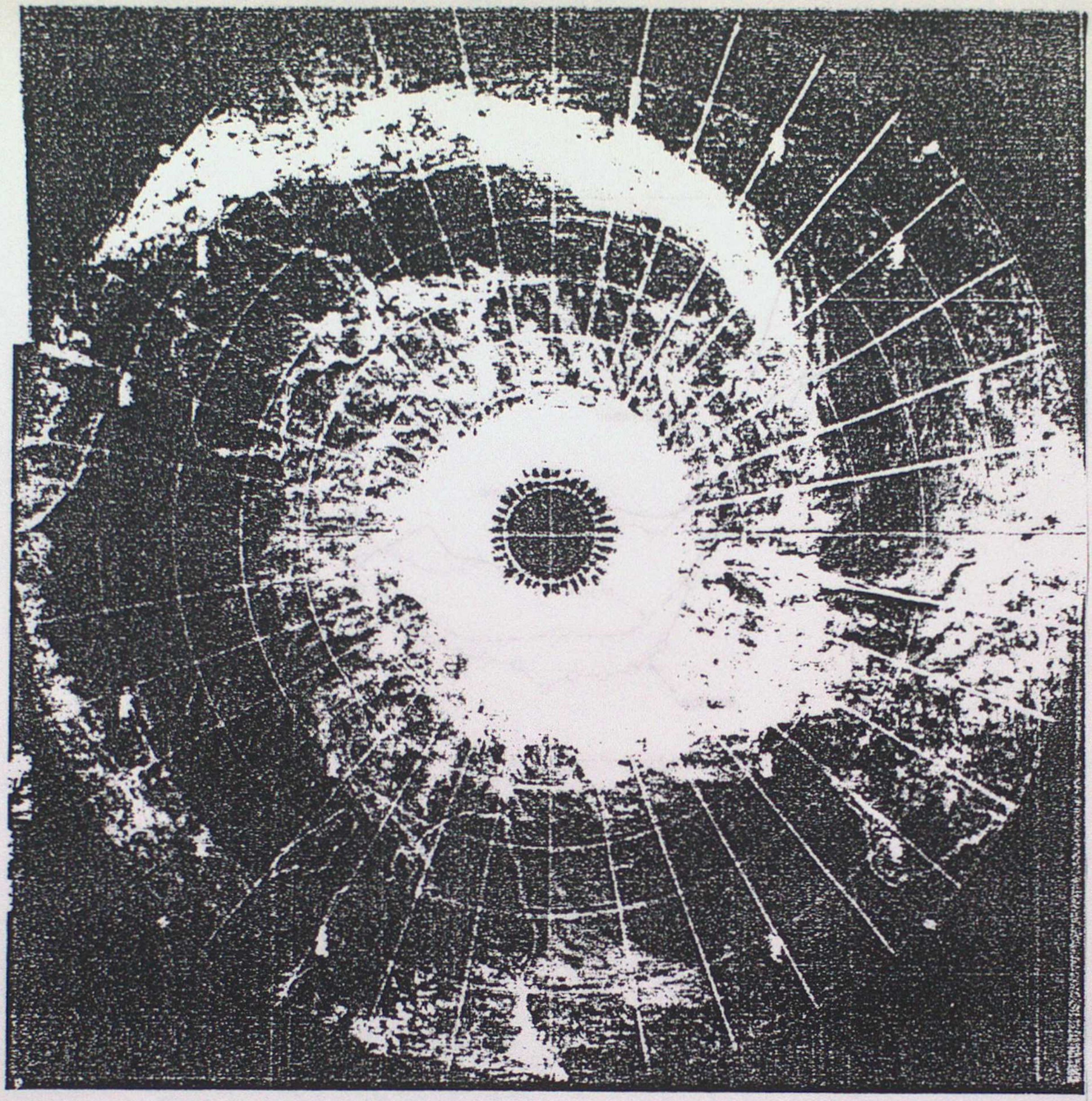
c) Average pressure (mb) at mean sea level, January 1974.
d) Average screen-level temperature anomaly (°C), January 1974. (Anomalies mainly from 1931–60 station averages.)

Figure 2.26 From Folland (1983)



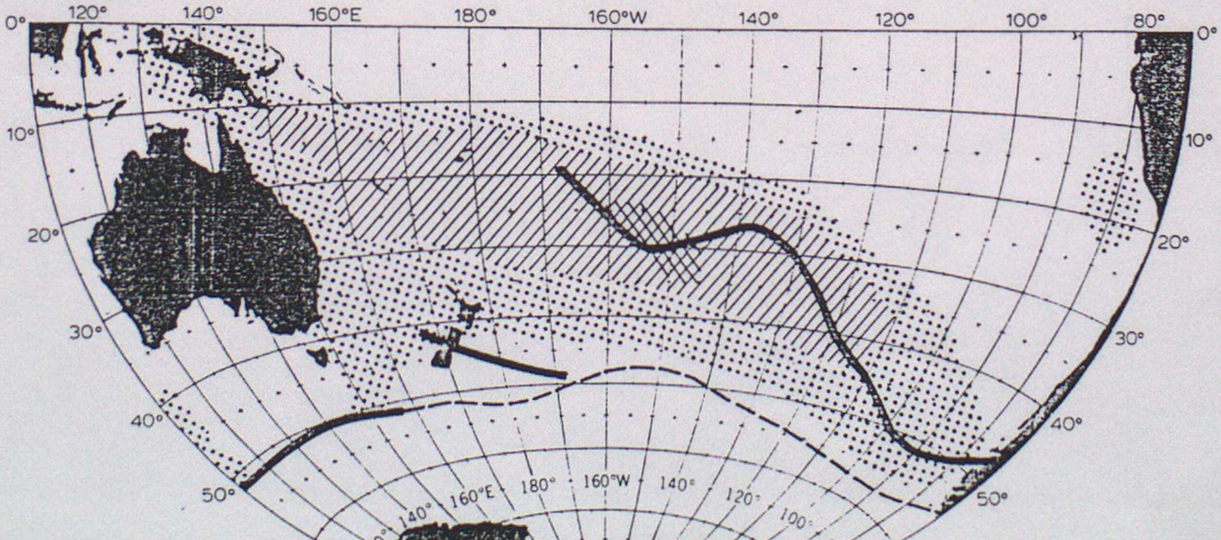
Vertical cross sections oriented perpendicular to a Southern Hemisphere low-latitude coastline (land to right of 0 km). Upper panel: Deviations of temperature °C (dashed lines) from the temperature at the coastline. A sea breeze circulation results (flow is parallel to solid lines, in sense of arrows). Lower panel: Solid lines are isotachs (lines of equal wind speed in m/s) showing a low level jet from the south (i.e., into the page). (*Courtesy of H. H. Lettau, "Explaining the World's Driest Climate," in H. H. Lettau and K. Lettau (eds.),* Exploring the World's Driest Climate, *Institute for Environmental Studies, Rep. 101, University of Wisconsin, Madison, Wis., 1978, pp. 182–248*)

Figure 2.27 From Trewartha and Horn (1980)



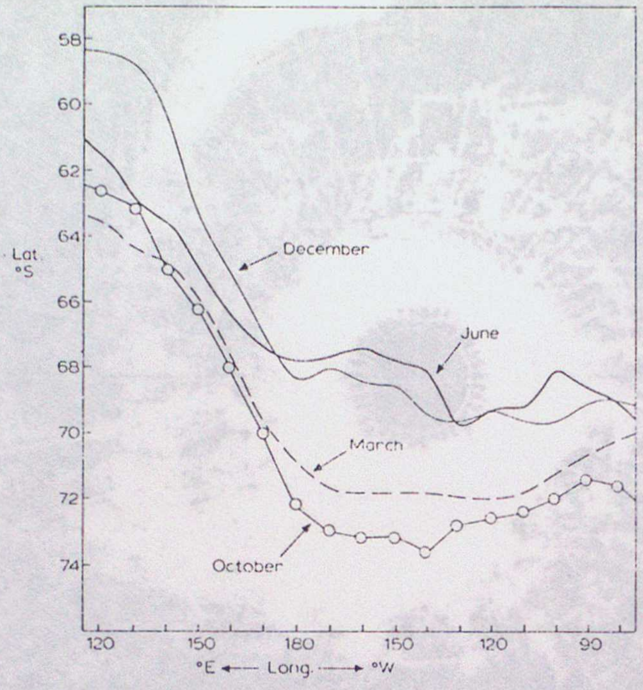
Averaged Southern Hemisphere cloud imagery (visible range) 1 – 15 October 1967, showing South Pacific cloud band. (After KORNFIELD and HASLER, 1969.)

Figure 2.28 From Streten and Zillman (1984)



Percentage frequency of 5-day averaged mosaics having axes of major cloud bands within a 5°-lat. by 10°-long. area for spring and autumn (stippled >10%; hatched >20%; cross-hatched >40%), after STRETEN (1973). Axes of zones of highest frequency of early cyclonic development for the same seasons is shown by the full line (after STRETEN and TROUP, 1973).

Figure 2.29 From Streten and Zillman (1984)



Mean latitude of the axis of the Antarctic trough in indicated months (1972 – 1977). (Data from STRETEN, 1980.)

Figure 2.30 From Streten and Zillman (1984)

Köppen's world-wide classification system

Climates are ranged in the following main groups:

A Tropical rain climates
B Arid climates
C (Warm) temperate rain climates
D Boreal forest and snow climates
E Cold snow climates (treeless)

A, C and D are all essentially tree climates; BS is the steppe grassland climate; BW defines the (warm) deserts (W standing for desert-waste/*Wüste*); E climates are those of the tundra and polar desert.

Subdivisions are designated as follows:

s Summer dry season
In group A climates confined to exceptional areas such as lee-sides of islands in the Trade Wind zone.
In groups C and D the wettest month of the year must have ⩾ 3 times the average rainfall of the driest month.

w Winter dry season
Group A climates are only designated Aw when at least one month normally has < 60 mm rainfall.
In groups C and D the wettest month of the year is required to have ⩾ 10 times the average rainfall of the driest month.

s′, w′ Special cases of summer or winter dry season climates, where the chief rains come in late summer–autumn.

x Climates where the early summer is wet, late summer fine.

s″, w″ Special cases of summer or winter dry season climates, where the rainy season is split – i.e. has two peaks, with another, short dry season in between.

f Moist climates, with no marked dry season.

m Monsoon climates, with only a short dry season in the winter half-year.

On this system the chief types of climate appearing on the world map (fig. 11.8) are defined as follows:

Tropical rain climates

Mean temperature of the coldest month in all cases > +18°C.

Af Driest month has on average ⩾ 60 mm rain.

Am Driest month < 60 mm but rainfall in the rainy season compensates this enough to allow forest.

Driest month R	60	40	20	0 mm
Required yearly R	1000	1500	2000	2500 mm

(Af and Am are both characteristically accompanied by tropical evergreen rain-forest.)

Aw Driest month < 60 mm: annual rainfall insufficient to compensate.
(Aw produces Savanna.)

Arid climates

Defined by (i) $R < 2T + 28$ where summer is the season with most rain.
or (ii) $R < 2T + 14$ where there is no season with more liability to rainfall than another.
or (iii) $R < 2T$ where winter is the season with most rain.

In these formulae R is the average rainfall for the year in centimetres; T is the average yearly temperature (°C).

BS $R \geqslant T + 14$ for summer rain areas
$R \geqslant T + 7$ where there is no season with more liability to rain than another
$R \geqslant T$ for winter rain areas
(In BS climates the vegetation ranges from bush to grassland.)

BW $R < T + 14$ for summer rain areas
$R < T + 7$ where there is no season with more liability to rainfall than another
$R < T$ for winter rain areas
(BW produces desert.)

Temperate rain climates

Mean temperature of the coldest month between +18°C and −3°C.

Cs Wettest (winter) month of the year has ⩾3 times the average rainfall of the driest (summer) month.
(Cs is typically accompanied by evergreen broad-leafed forest.)

Cf No marked dry season.
Sub-types:
Cfa Hot summers: warmest month > + 22°C
Cfb Warm summers: at least 4 months > + 10°C
Cfc Cool summers: only 1–3 months > + 10°C.
(Cfa and Cfb vegetation generally deciduous broad-leafed forest, but in areas with very mild winters evergreen broad-leafed forest.
Cfc vegetation needle-tree forest.)

Cw Wettest (summer) month has ⩾10 times the rainfall of the driest winter month.
(Cw, occurring mainly on mountain heights above Aw climates, produces evergreen forest vegetation.)

Boreal forest and snow climates

Mean temperature of the warmest month > + 10°C, coldest month < − 3°C.

Dw Wettest (summer) month has ⩾10 times the average precipitation of the driest (winter) month.

Df Less seasonal difference in the average downput of rain and snow.
Sub-types:
Dfa and Dwa Hot summers: warmest month > + 22°C
Dfb and Dwb Warm summers: at least 4 months > + 10°C
Dfc and Dwc Cool summers: only 1–3 months > + 10°C.
(Dfa, Dfb, Dwa, Dwb climates go with deciduous (broad-leafed) forest.
Dfc, Dwc go with needle-tree forest.)

D climates and their vegetation are only found in the northern hemisphere, where the great continents provide the requisite conditions.

Cold snow climates

Mean temperature of the warmest month < + 10°C.

ET Mean temperature of the warmest month > 0°C.
(ET vegetation is tundra, dwarf tree species and mosses.)

EF (sometimes called simply F).
Mean temperature of the warmest month <0°C.
Brief thaws and rain can occur, but have little or no effect on the long-term condition of the surface.
(No vegetation in EF climates: polar desert.)

Further letters are sometimes used to distinguish particular climates and subdivisions of the KÖPPEN types:

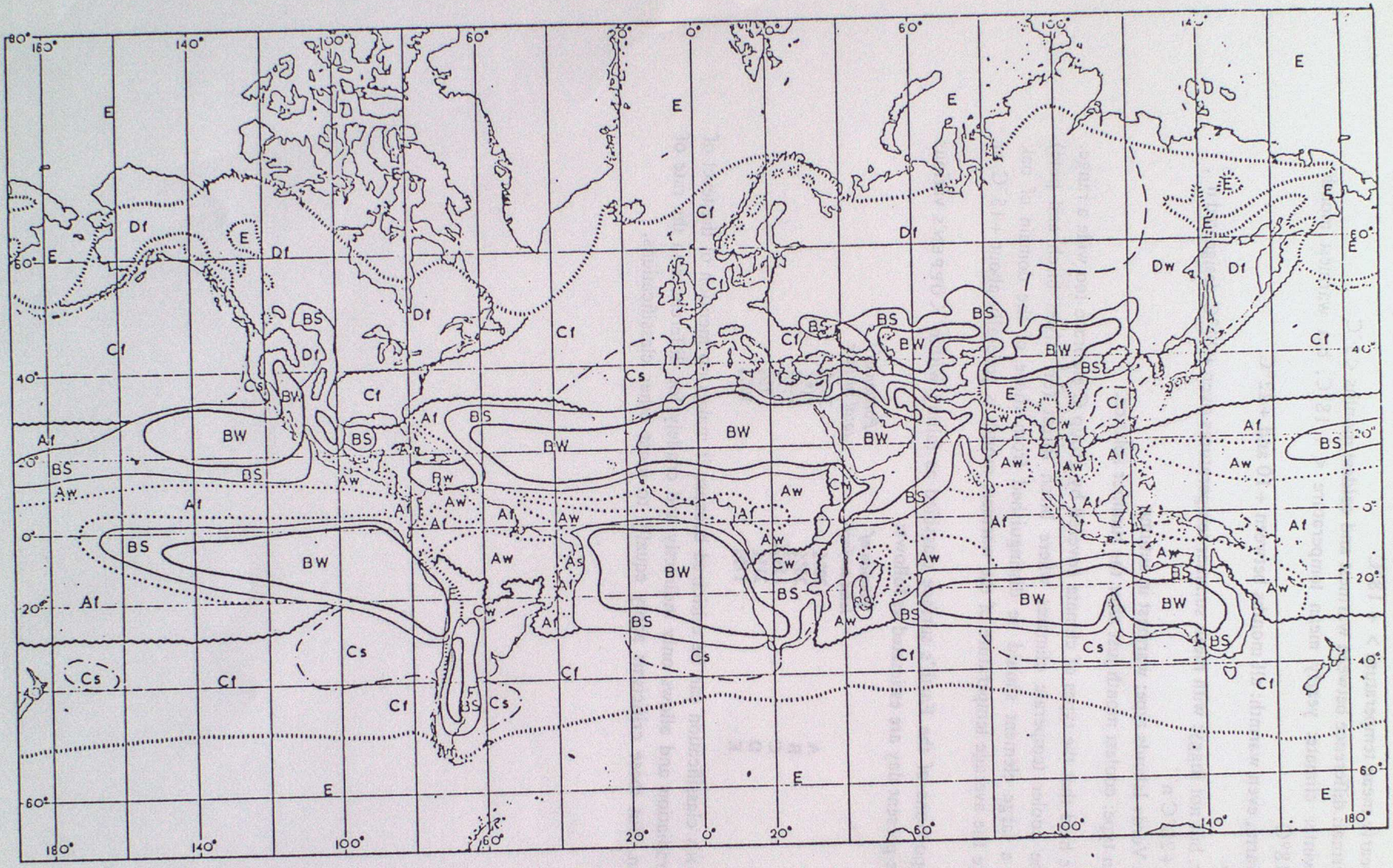
d	coldest month < − 38°C
g	Ganges type climate, warmest month before the summer solstice, height of summer rainy
h	hot: yearly mean temperature > + 18°C
i	isothermal: difference between warmest and coldest months < 5°C
k	cold-winter climate: yearly mean temperature < + 18°C, but warmest month > + 18°C
l	lukewarm, even warmth: all months between + 10 and + 22°C
n	foggy
n′	moist, but not foggy: with mean summer temperatures over + 24°C designated **n″**, over + 28°C **n‴**
t′	Cape Verde Islands type: warmest in autumn
t″	Sudan type: coolest month just after the summer solstice.

It may be held that the types of climate covered by group C embrace too wide a range, and that the cooler temperate climates where birch forest (or mixed birch and pine) constitutes a large element should be distinguished from those of the domain of oak forest where the average temperature of the warmest month is generally about + 15°C or over.

The proportions of the Earth's surface classified as falling within KÖPPEN's various types at the present day are estimated as follows:

	Portion of land surface	*Portion of total surface*
A	20%	36%
B	26%	11%
C	15½%	27%
D	21%	7%
E	17%	19%

KÖPPEN's classification can be criticized because it makes no mention of the level of incoming radiation and allows only indirectly and crudely for differences in the rate of evaporation. But these criticisms apply equally to most other classifications.



A-C boundaries
BW boundaries
BS boundaries
C-D boundaries
E boundaries
internal boundaries between A climates
internal boundaries between C climates
internal boundaries between D climates

Figure 2A.1 KÖPPEN's world classification of climates.

After Lamb (1972)