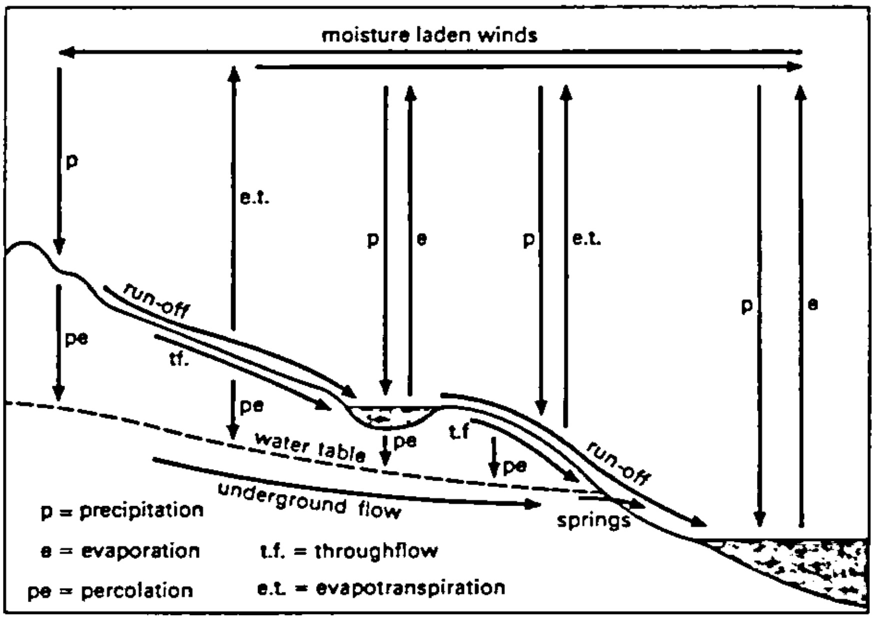
AGMET

The Balance of Water - Present and Future



Proceedings of

AGMET Group (Ireland)
and Agricultural Group of the Royal
Meteorological Society (UK)

Conference

Trinity College Dublin, September 7 - 9, 1994

General Editors: T. Keane and E. Daly

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Foreword

Considerable reliance is placed on climatological and hydrological data for a wide range of water balance studies. The accuracy and archival status of the input data frequently concern hydrologists and agriculturals, who also ask what confidence can be attached to estimates of the different components. Apart from the reliability of instrumentation, the limitations of point measurement or estimation by formula, further difficulties arise from the specification of soil characteristics, effects of farming practices, distance from sea, altitude, and indeed from the method of spatialization of the data. This Dublin Conference sets out to review the research and progress of recent decades, address some of the current issues and indicate the likelihood of more accurate specification and determination of the parameters: precipitation, evaporation/ evapotranspiration and run-off. The measurement and application of the water balance is central to the study of hydrology. A breath of scholarship together with the most recent research are brought to bear in the following papers presented at the Conference.

The AGMET Group, was founded in 1984 to help develop and facilitate the practical application of meteorological services to Irish agriculture and farming. It is composed of experts drawn from the Meteorological Service, Teagasc (Agriculture and Food Development Authority), Geological Survey of Ireland, Office of Public Works, Veterinary Services, and university departments concerned with agriculture, hydrology and the environment. This is the third major conference to be undertaken by the Group. The UK Agricultural Group is one of a number of specialist groups of the Royal Meteorological Society. In the furtherance of the science of agrometeorology it regularly undertakes conferences in Britain on a wide range of related topics. This Conference is the fruit of a unique joint venture between both organizations.

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Acknowledgments

Sincerest thanks is expressed to Conference speakers and poster presenters, their parent institutes and the supporting organizations who helped to make this Meeting possible. Bórd na Móna, Dr Eugene Bolton and Ms Catherine Meenan are especially thanked for the visit arrangements to their Research Laboratories in Newbridge and for Corporate hospitality. The Organizing Committee wishes to thank the secretaries in the various institutes who typed authors' papers onto disc to a specified format. The continued support of the AGMET Group members has been essential to the success of the Conference. The sustained support of Professor Jim Dooge, who is Chairman of the Conference, is gratefully acknowledged and appreciated. A special thanks to Dr Mike Jones and Ms Pamella Hennessy, Botany Department, Trinity College Dublin, for providing the Secretariat to the Conference. Sincere thanks Dr Jerry Clark, Secretary of the Agricultural Group of the Royal Meteorological Society, for his central role in promoting this joint meeting in Britain.

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Programme

Tuesday, September 6

Afternoon and evening: arrival of delegates and poster mounting

18.00 Conference registration, Luce Hall
20.00 Reception in Luce Hall, Trinity College Dublin
Welcome by Mr Declan Murphy, Director, Meteorological Service; and
Mr Liam McCumiskey, Director General, Environmental Protection Agency

Hamilton Building
Theatre 4

Wednesday, September 7

09.00 Registration
09.30 Opening of Conference by Mr John Browne TD, Minister for Environmental Protection

Conference Chairman: Professor Jim Dooge

The Measurement of the Water Balance

Morning

Chair : **Jim Dooge**

09.45 Weighing up the Water Balance
John C Rodda, World Meteorological Organization, Genèva
10.30 Discussion

10.45 Coffee

11.00 Fifty Years of Potential Evaporation
John L Monteith, Institute of Terrestrial Ecology, Edinburgh
11.45 Discussion

Chair : **Mike Jones**

12.00 Poster/Short Presentations

13.00 Lunch

Afternoon

Chair : **John Rodda**

14.30 Adequacy of the Irish Rainfall Network
Denis L. Fitzgerald, Meteorological Service, Dublin

15.10 River flow in Ireland - Characteristics and Measurement
John V Martin, Office of Public Works, Dublin and Con Cunnane, University College Galway

15.50 Discussion

16.05 Coffee

Chair : **Reinder Feddes**

16.25 Similarity and Scale Effects in the Water Balance of Heterogeneous Areas
Keith J Beven, CRES, Lancaster University

17.05 Discussion

Chair : **Mike Jones**

17.20 Poster overview

18.00 End of Session

Thursday, September 8

09.00 Depart Dublin for Newbridge, Co Kildare

10.15 Bórd na Móna research facilities and corporate presentation
Dr Eugene Bolton, Director of Research

Aspects of Blanket Bog Research at Glenamoy
William(Billy) Burke, formerly An Foras Talúntais, Dublin

The Hydrodynamics of Raised Bogs: An Issue for Conservation
Donal Daly, Geological Survey of Ireland, Dublin and Paul Johnston, Trinity College Dublin

12.15 Lunch at Bórd na Móna

13.00 Tour through midland bogs (active and preserved bog/fen)
Clara bog - ecology, hydrology, conservation

15.30 Coffee at Clara

19.00 Civic Reception at Mansion House hosted by Rt. Hon. The Lord Mayor, Councillor John Gormley

Friday, September 9

APPLICATIONS OF WATER BALANCE

Chair : John Monteith

09.30 **Water Balance of Wetland Areas**
Kevin Gilman, Institute of Hydrology, Plynlimon

10.05 **Broadleaved Woodland, its Water Use and some Aspects of Water Quality**
Robin Hall, Institute of Hydrology, Wallingford

10.40 **Discussion**

11.00 Coffee

Chair : Con Cunnane

11.25 **Modelling Catchment-scale Water Balance Dynamics using Long Time Series of Rainfall, Stream-flow and Air Temperature**
Ian G Littlewood, Institute of Hydrology, Wallingford

12.00 **Spatial Variability and Accuracy of Remote Sensing Estimates of Evaporation and Soil Moisture within EFEDA '91**
Reinder A. Feddes, Wageningen Agricultural University, The Netherlands

12.40 **Discussion**

13.00 Lunch

Chair : Keith Beven

14.15 **Water : Some Relationships between Quantity and Quality**
Marie Sherwood, Environmental Protection Agency, Wexford

14.45 **Hydrological and Landuse Changes in North Antrim**
Nicholas L Betts, The Queen's University of Belfast

15.15 **Climate Change and a New Water Balance**
James C I Dooge, CWRR, University College Dublin

15.45 **Discussion**

16.15 **Close of Conference**

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THE MEASUREMENT
OF THE
WATER BALANCE

WEIGHING UP THE WATER BALANCE

John C. Rodda
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Abstract

The history and development of water balance studies are outlined including basin studies and large scale experiments. Then comments are presented on the nature of the instrument networks which furnish the data for determining balances and the errors inherent in them. The work of WMO to reduce these errors is discussed and the concept of total quality management (TQM) is introduced. The conclusion reached is that without TQM, it will remain difficult to weigh up the water balance.

1. Background

Water has been important to mankind since the start of history. However, concepts concerned with its distribution about the globe have only been turned into facts during the last 300 years. Now the theories from ancient Greece, Rome and the Renaissance on the hydrological cycle and its pathways seem strangely complicated, particularly by the mechanisms needed in the subterranean route to raise water to the surface of the ground. But with the birth of scientific hydrology in 1674 (UNESCO/WMO/IAHS 1974), following the publication of "De l'origine des fontaines" by Pierre Perrault (Perrault 1674), there was the first experimental evidence that atmospheric transport provides precipitation in sufficient quantities to cause rivers to flow. Perrault's study of the Coquille at Aignay le Duc, a headwater tributary of the Seine, and the contemporary work of Mariotte, on the basin of the Seine to Paris (Tixeront 1974), (Table 1) are the earliest assessments of the water balance of any basin. Doubts that lingered during the remainder of the 18th Century over the ability of the atmosphere to transport sufficient moisture were finally scotched by Dalton (1802). He combined lysimeter-made measurements of evaporation with discharge estimates based on measurements carried out on the Thames and with a large number of records of rainfall, to estimate the water balance for England and Wales.

TABLE 1
Seventeeenth Century water budgets from the Perrault and Mariotte studies in the Seine basin (from Tixeront 1974)

	Perrault - Aignay-le-Duc		Mariotte - Paris	
	D'après l'auteur	Données réelles actuelles	D'après l'auteur	Données réelles actuelles
Surface du bassin (km²)	118,8	93	53500	44320
P = pluie annuelle (mm)	518	900	459	750
R = écoulement annuel:				
millions de m³	9,5	31,5	3574	8600
mm par an	80	340	67	194
Déficit : P - R (mm)	438	560	392	556

In the industrializing regions of the world, the 19th Century saw an increase in the ability to measure the different hydrological variables (Table 2) and a gradual expansion of countrywide instrument networks. It also saw the development of the skills of collecting and analysing hydrometeorological data and the organization of the earliest national hydrological and meteorological services. Some of these analyses were for purely scientific purposes, but an increasing number were aimed at the design and operation of various structures and schemes to cope with the increasing demand for water, to provide drainage and to alleviate floods. Water balance studies were important to a number of these purposes and to questions which began to be asked about the impact of human activities on the water balance. These studies involved the range of scales (Dooge 1984) pertinent to water as a resource; at one extreme, studies of the global water balance, at the other studies of experimental plots and small basins.

TABLE 2
Most widely measured hydrological variables

Variable	Unit
Precipitation (rain, snow)	depth in mm and cm (coverage in % and density of snow)
Water level in rivers (and the flows determined from them), lakes and reservoirs	depth in m and cm (flow in m³/s)
Water level in wells and boreholes	depth in m and cm
Evaporation and evapotranspiration	depth in mm
Soil water content	% volume or % mass
Sediment concentration in rivers (and the loads determined from them)	kg/m³
Water quality concentration in rivers (and the loads determined from them), lakes, reservoirs and boreholes	mg/litre (tonnes)
Ice depth and cover	cm and %

2. Approaching the balance

One of the earliest examples of a water balance study on the smaller scale commenced in the 1890s in the basins of the Sperbelgraben and Rappengraben in the Emmental, Switzerland (Engler 1919). This investigation into the hydrological differences between forest and pasture was the first in a long series of paired basin studies, such as Waggon Wheel Gap (USA), Coweeta (USA), Valdai (Russia), Hupsalse Beek (Netherlands), and Plynlimon (UK) (Rodda 1976, Robinson & Whitehead 1993) aimed primarily at determining the hydrological impact of land use differences and changes. These studies were stimulated by UNESCO's International Hydrological Decade (IHD) and its subsequent International Hydrological Programme (IHP), a number of the results being employed to assist in the development of the comparative approach to hydrology (Falkenmark and Chapman 1989). More recently, these basin studies and the application of the results from them have been promoted in Europe by the IHP Frend Project (Roald et al 1989) and in other parts of the world by the series of similar projects the Frend concept has fostered. Utilizing the regional data bases it develops, Frend has two main objectives. The first is the application of these bases to water resources problems, such as the estimation of extreme events for sites with no records. The second is assessment regionally of the impact of human activities on the hydrological cycle.

The first studies of the world water balance also seem to have commenced in the late nineteenth century: a series of examples being listed by Lvovitch (1970 and 1979) and Baumgartner and Richel (1970). More recently Korzun (1974) and Shiklomanov (1991) have assessed in detail the global budget and its regional patterns, while the variations of its components have been summarized on a number of occasions, such as by Street-Perrott et al (1983). Table 3 sets out the present state of knowledge of the world water balance, drawing on a number of sources, while Fig. 1 portrays the balance diagrammatically. In recent years, knowledge of the processes that determine the water balance has been improving because of process studies within basin studies and because of the series of large scale experiments on land/atmosphere interactions being conducted in different parts of the world (Shuttleworth, 1988 and 1991). Now some of these experiments come under the International Geosphere Biosphere Programme (IGBP) core project on the Biological Aspects of the Hydrological Cycle (Bolle et al, 1993). Others fall within the World Climate Research Programme, Global Energy and Water Cycle Experiment (GEWEX) (ICSU/WMO 1990) and its Continental Scale International Project (ICSU/WMO, 1992), focussing on the Mississippi Basin.

Because the hydrological cycle provides much of the power to transport materials in the different geochemical cycles, measurements of the fluxes in the water balance globally and basin-wide are essential in determining the budgets of these different materials. Consequently, knowledge of runoff is needed to estimate the transported loads of sediment (Walling & Webb 1987), carbon and other determinands (Degens et al 1984, Meybeck 1982); Table 4 shows estimates of the transport of material in suspension and solution from the land mass to the world ocean. In addition, precipitation amounts must be known in order to determine the loads being deposited from the atmosphere (Goodison & Vet 1989). A summary of the movement of water and material about the globe has been made by Berner and Berner (1987).

TABLE 3

Approximate quantities of water in the various parts of the hydrological cycle with replacement periods

Category	Total volume ($km^3 \cdot 10^3$)	% of total	% of fresh	Annual volume recycled (km^3)	Replacement period
Oceans	1338000	96.5		505000	2654 y
Groundwater to 2000m	23400	1.7		16700	1400 y
Predominantly fresh groundwater	10530	0.76	30.1	—	—
Soil Moisture	16.5	0.001	0.005	16500	1 y
Glaciers and permanent snow	24064.1	1.74	68.7	—	—
Antarctica	21600	1.56	61.7	—	—
Greenland	2340	0.17	6.68	2477	9700 y
Arctic Islands	83.5	0.006	0.24	—	—
Other Mountain areas	40.6	0.003	0.12	25	1600 y
Ground ice (permafrost)	300	0.022	0.86	30	10000 y
Lakes	176.4	0.013		10376	17 y
Freshwater lakes	91	0.007	0.26	—	—
Salt water lakes	85.4	0.006		—	—
Marshes	11.47	0.0008	0.03	2294	5 y
Rivers	2.12	0.0002	0.006	49400	16 d
Biological water	1.12	0.0001	0.003	—	—
Atmospheric water	12.9	0.001	0.04	600000	8 d
Total water	1385984.61	100*			
Total freshwater	35029.21	2.53	100*		

* Some duplication in categories and sub-categories.

After Kalinen and Bykov (1969), Korzun (1974), L'vovitch (1974) and Nace (1969).

From Young, Dooge and Rodda (1994).

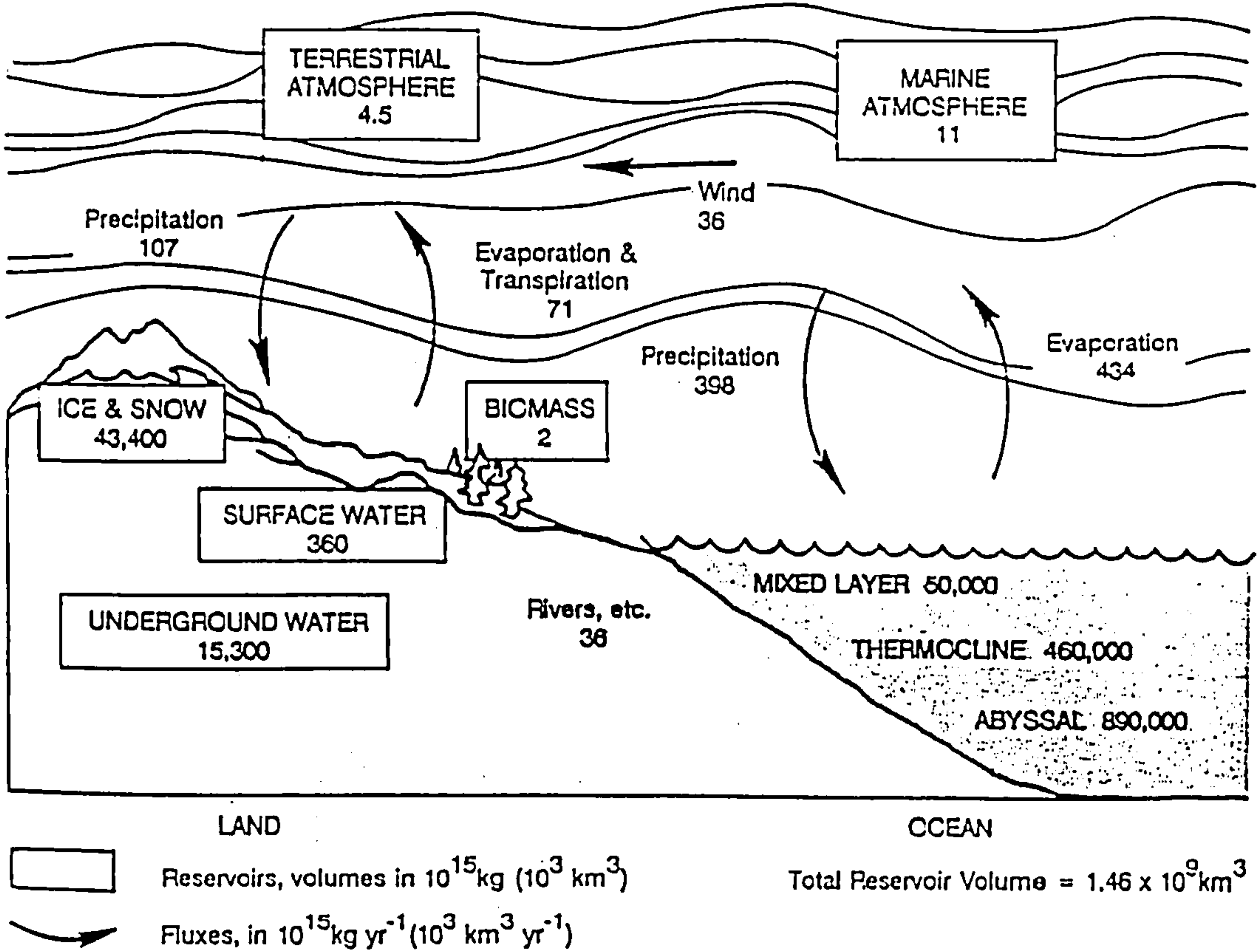

Figure 1 - The hydrological cycle at global scale (from Opportunities in the Hydrologic Sciences, National Research Council 1991)

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TABLE 4

Material transport from the continents to the oceans (based on load estimates produced by Milliman & Meade (1983) and Meybeck (1979) from Walling & Webb (1987))

CONTINENT	SUSPENDED SEDIMENT		DISSOLVED		SEDIMENT/ DISSOLVED RATIO
	(10^6t year^{-1})	(t km^{-2}year^{-1})	(10^6t year^{-1})	(t km^{-2}year^{-1})	
Africa	530	35	201	13	2.6
Asia*	6433	229	1592	57	4.0
Europe	230	50	425	92	0.5
North and Central America	1462	84	758	43	1.9
Oceania/Pacific Islands+	3062	589	293	56	10.5
South America	1788	100	603	34	3.0

*Mainland Asia, includes Eurasian Arctic.
+Includes Australia and the large Pacific Islands.

3. Instrument networks

Of course, data on the water balance at the global scale, or for the smallest headwater basin, must be determined by measurements. Traditionally these measurements have been derived from networks of ground-based instruments, but now data are available in an increasing volume from weather radars and from satellite imagery. However, in only a few countries are these data used routinely in assessments of the water balance.

Table 5 provides a summary of the statistics for the global instrument network, compiled from statistics on national networks, which in some cases, include the networks employed in basin studies (WMO 1994a). These instruments and methods of observation are operated on a routine basis by the world's Hydrological and Meteorological Services who collect, analyze and apply the data from them. Figures 2 and 3 show the contrasting densities for parts of this network. It is very evident that for many parts of the world and for certain variables, coverage is poor. For other parts, the networks are dense and most of the variables listed in Table 2 are measured, while much of the data produced appear to have the desirable characteristics of reliability, continuity and representativity. However, the contrasts between the data rich and data poor areas of the world, and particularly the fact that 70% of the globe lacks measurements of precipitation and evaporation, rarely feature as commentaries on published global water budgets. Error, accuracy and precision are words that seem to be absent from most of these discussions: however, they appear more frequently in the results of small basin studies. They are, of course, the concern of those who operate these instrument networks and manage the data obtained from them on a regular basis, as they strive to maintain the quality of the data in the hydrological information system (Fig. 4). Agency-wide, national and international programmes aid this effort to assure quality. Certain initiatives on the international level aim to assist in quality assurance, as well as in making international data sets more readily available. There are, for example, the World Glacier Monitoring Service in Zurich, the Global Runoff Data Centre in Koblenz, the Global Precipitation Climatology Centre at Offenbach and the Collaborating Centre for Surface and Groundwater Quality at Burlington. Each of these centres quality controls the data it acquires before archiving them and, in addition, as part of the GEMS Water Quality Programme (WHO 1991), a considerable amount of help is provided to national services to improve and maintain the standards of their analytical services for water quality. Unfortunately, the data held by these centres does not cover all countries, the most recent are frequently two or three years old and time series are often incomplete.

TABLE 5 - HYDROLOGICAL OBSERVING STATIONS - SUMMARY*
(Number of Stations)

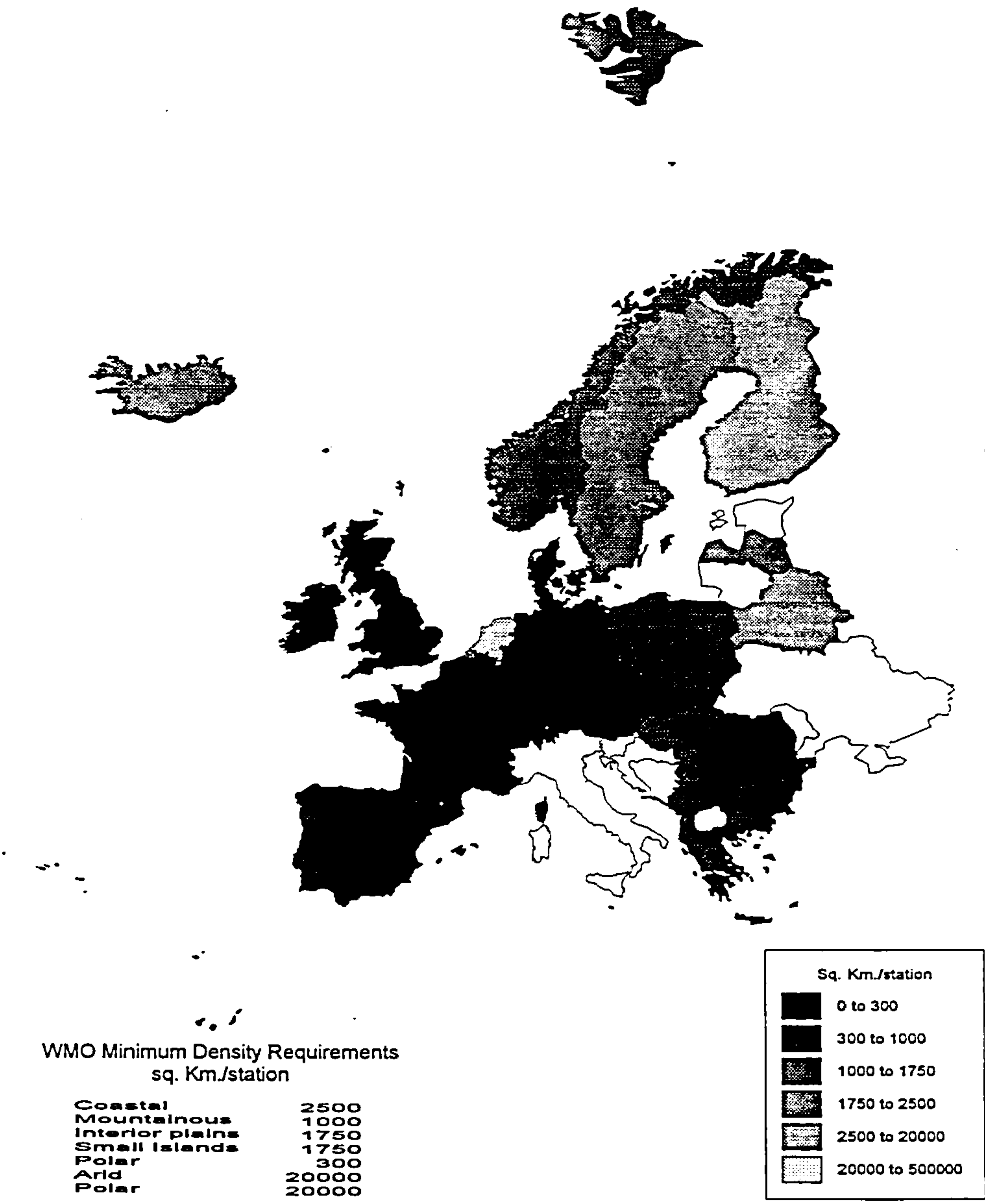
TYPE OF STATIONS	WMO REGIONS: AFRICA (RA I) (1)	ASIA (RA II) (2)	S. AMERICA (RA III) (3)	N. & C. AMERICA (RA IV) (4)	S.W. PACIFIC (RA V) (5)	EUROPE (RA VI) (6)	TOTAL (GLOBAL) (7)
PRECIPITATION							
Non-recording: TOTAL ***	17 036	39 456	19 247	19 973	15 276	40 367	151 355
(0- 500m)	1 304	11 478	8 548	5 923	12 881	28 681	68 815
(501-1000m)	1 206	2 076	3 386	1 196	1 463	7 083	16 410
(1001-1500m)	1 886	1 314	663	428	403	2 234	6 928
(1501-2000m)	339	1 047	422	130	199	530	2 667
(2001-2500m)	29	641	256	38	29	143	1 136
(over 2500m)	8	653	626	16	0	94	1 397
Recording: TOTAL ***	2 639	18 864	4 124	5 280	3 332	8 422	42 661
(0- 500m)	277	5 168	2 330	1 205	2 700	6 233	17 913
(501-1000m)	414	1 168	757	269	389	1 609	4 606
(1001-1500m)	780	385	229	129	118	436	2 077
(1501-2000m)	92	194	133	49	21	109	598
(2001-2500m)	9	84	91	23	4	9	220
(over 2500m)	1	17	100	10	0	3	131
Telemetry	8	1 916	211	1 023	515	459	4 132
Radar	9	56	3	82	8	35	193
EVAPORATION							
Pans	1 508	3 686	2 031	2 716	1 120	1 499	12 560
Indirect method	374	7	40	11	1 049	488	1 969
DISCHARGE: TOTAL **	5 703	11 543	7 924	13 211	5 838	19 798	64 017
Recording	1 856	3 064	2 233	11 128	3 795	13 661	35 737
Non-recording	3 045	8 479	5 691	2 080	2 043	6 137	27 475
Telemetry	39	2 033	158	3 613	1 075	2 561	9 479
STAGE (WATER LEVEL): TOTAL **	3 410	6 405	5 872	11 274	1 167	10 474	38 602
Recording	877	2 300	1 628	9 549	642	4 599	19 595
Non-recording	2 244	3 800	4 244	1 725	522	5 826	18 361
Telemetry	15	1 257	194	1 734	192	1 768	5 160
SEDIMENT DISCHARGE							
Suspended	859	3 820	1 561	5 217	619	3 712	15 788
Bedload	6	685	505	0	1	549	1 746

TABLE 5 - HYDROLOGICAL OBSERVING STATIONS - SUMMARY*
(Number of Stations)

TYPE OF STATIONS	AFRICA (RA I) (1)	ASIA (RA II) (2)	S. AMERICA (RA III) (3)	N. & C. AMERICA (RA IV) (4)	S.W. PACIFIC (RA V) (5)	EUROPE (RA VI) (6)	TOTAL (GLOBAL) (7)
	WMO REGIONS						
(continued . . .)							
WATER QUALITY	5 297	5 045	2 752	31 462	1 690	55 379	101 625
GROUNDWATER							
Water Level							
- Observation wells	4 884	16 657	1 133	19 818	18 585	85 075	146 152
- Production wells	31 804	63 705	14 150	14 099	13 504	38 452	175 714
Temperature							
- Observation wells	287	2 541	5 200	21 097	4 888	18 967	52 980
- Production wells	243	88	5 539	21 501	888	1 641	29 900
Water Quality							
- Observation wells .	4 898	1 964	320	13 757	7 935	14 889	43 763
- Production wells	5 674	45 187	3 416	14 825	3 172	23 711	95 985

* Summary, by Regions, of information given in Tables 4.1.01 to 4.6.12. Updated May 1994
** The total includes stations not distinguished as "recording" and "non-recording"
*** These totals do not correspond to the sums of stations located at different elevations, as not all countries provided elevation data

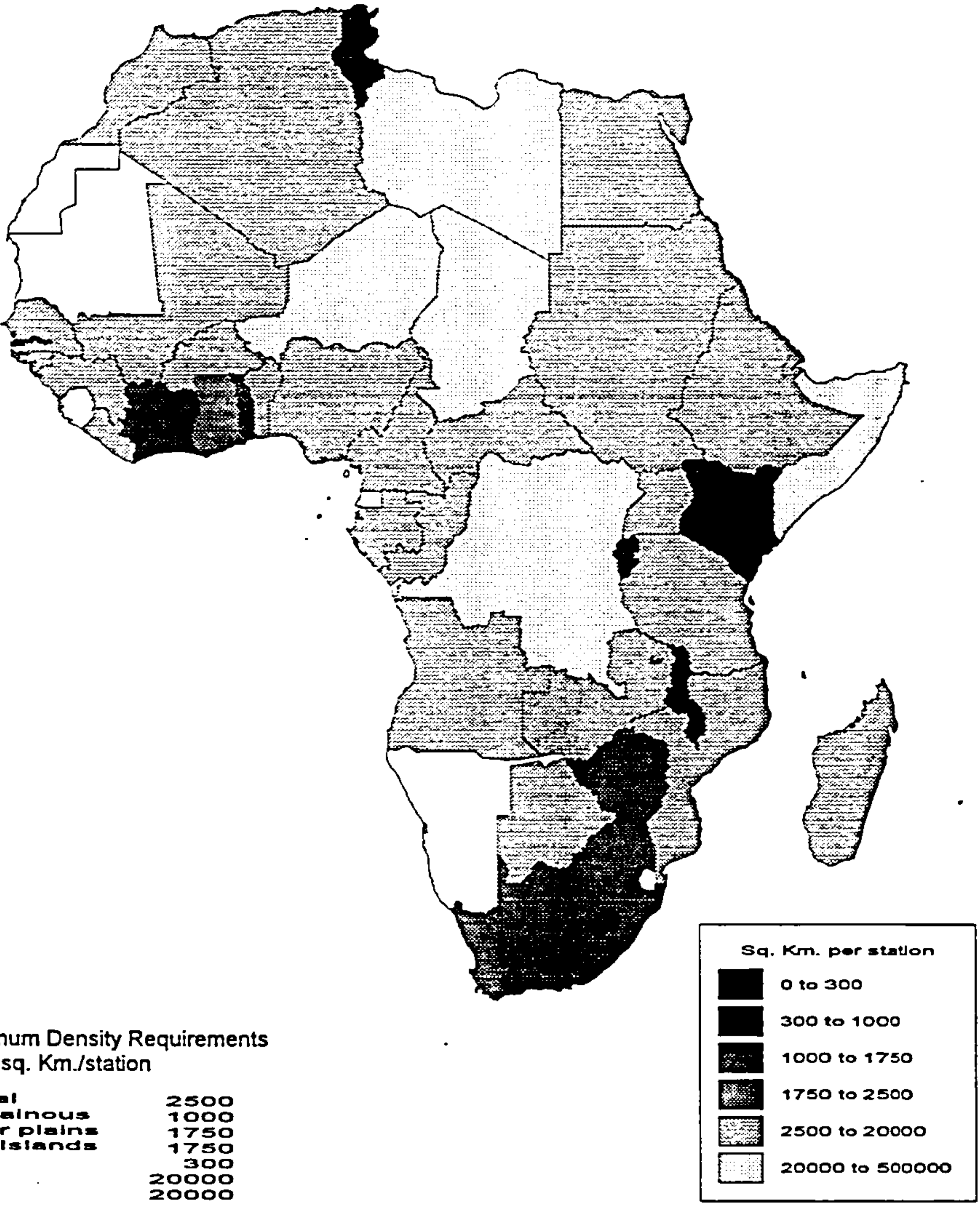
Figure 2. Network Densities in Europe - Discharge Stations



WMO Minimum Density Requirements
sq. Km./station

Coastal	2500
Mountainous	1000
Interior plains	1750
Small islands	1750
Polar	300
Arid	20000
Polar	20000

Figure 3. Network Densities in Africa - Discharge Stations



WMO Minimum Density Requirements
sq. Km./station

Coastal	2500
Mountainous	1000
Interior plains	1750
Small Islands	1750
Polar	300
Arid	20000
Polar	20000

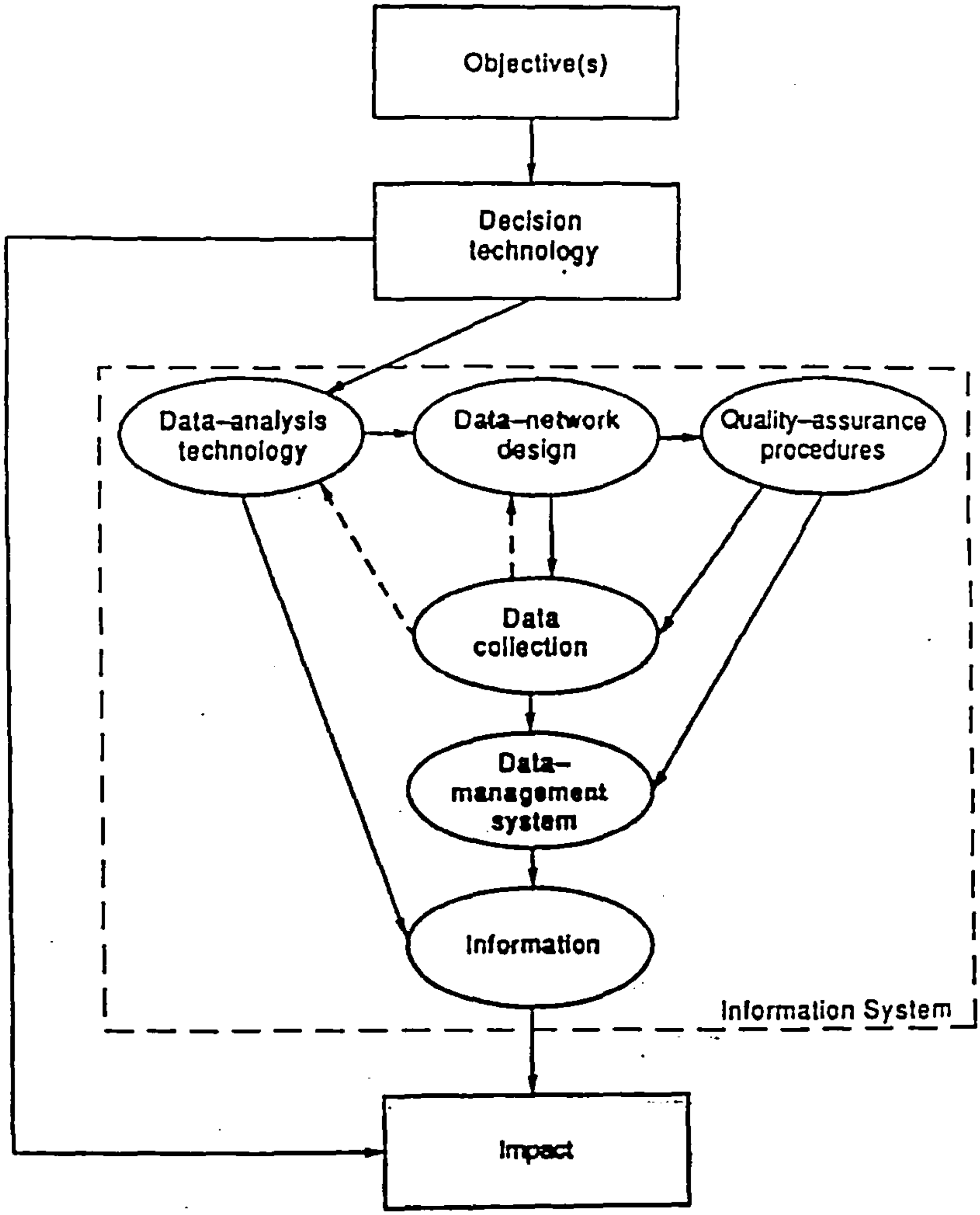


Figure 4 - Components of an idealised hydrological-information system (from 5th Edition, Guide to Hydrological Practices)

The absence of a readily accessible and reliable body of hydrological data for the globe has led WMO to propose the establishment of a World Hydrological Cycle Observing System (WHYCOS), which would consist of about 1,000 stations world-wide, sited on the major rivers (Rodda et al 1993). Each station would monitor about 15 variables, including flow and physico-chemical determinands of water quality, which would be transmitted via a geostationary satellite, such as Meteosat, to national, regional and global centres. At these centres, archives of data would be built up over the period of operation of WHYCOS (about 20 years) and processed to create tools for decision making, as well as for science. These archives could be extensions to existing national archives and to those compiled for Frend purposes, as well as to the existing global data centres. WHYCOS would also contribute to the Global Climate Observing System (GCOS), and the Global Terrestrial Observing System (GTOS), initiatives which WMO is sharing with a number of international bodies concerned with identifying global change. In addition, WHYCOS would also seek to build up the capabilities of the hydrological services in those countries where networks, staff levels and facilities are in decline. Such a decline has been revealed in a number of recent studies, a decline which is most marked in Africa (WMO/UNESCO, 1991, UNDP/World Bank, 1990-94). With this decline, an increase in the errors surrounding water balance estimates are to be expected, errors which would apply to most scales, basin-wide to global. Indeed, it is something of an enigma that at the time when global demand for water is rising faster than ever before, the errors in assessing just how much water is available for use are generally increasing.

For research on the water balance of small basins and for the process studies within them, the instrument systems are normally more advanced and more complete than for countrywide networks. Toebes and Ouryvaev (1970) provided an overview of observational and other practices for representative and experimental basins and there are many more recent reviews of experience in individual basin studies, such as IAHS (1980), Swanson et al (1987) and Lang & Musy (1990). The approach to basin studies is typified by the report for the Plynlimon catchments (Kirby et al 1991), Fig. 5 portraying the instrument network. Calder (1992) discusses the general design of basin studies and process studies and summarizes the results that have been drawn from them.

Large-scale studies of hydrological processes such as FIFE, HAPEX-MOBILHY and BOREAS have a much shorter history than basin studies and studies of the global water balance. Essentially, they couple simultaneous measurements of a number of variables on different scales in intensive field campaigns, measurements that have previously been

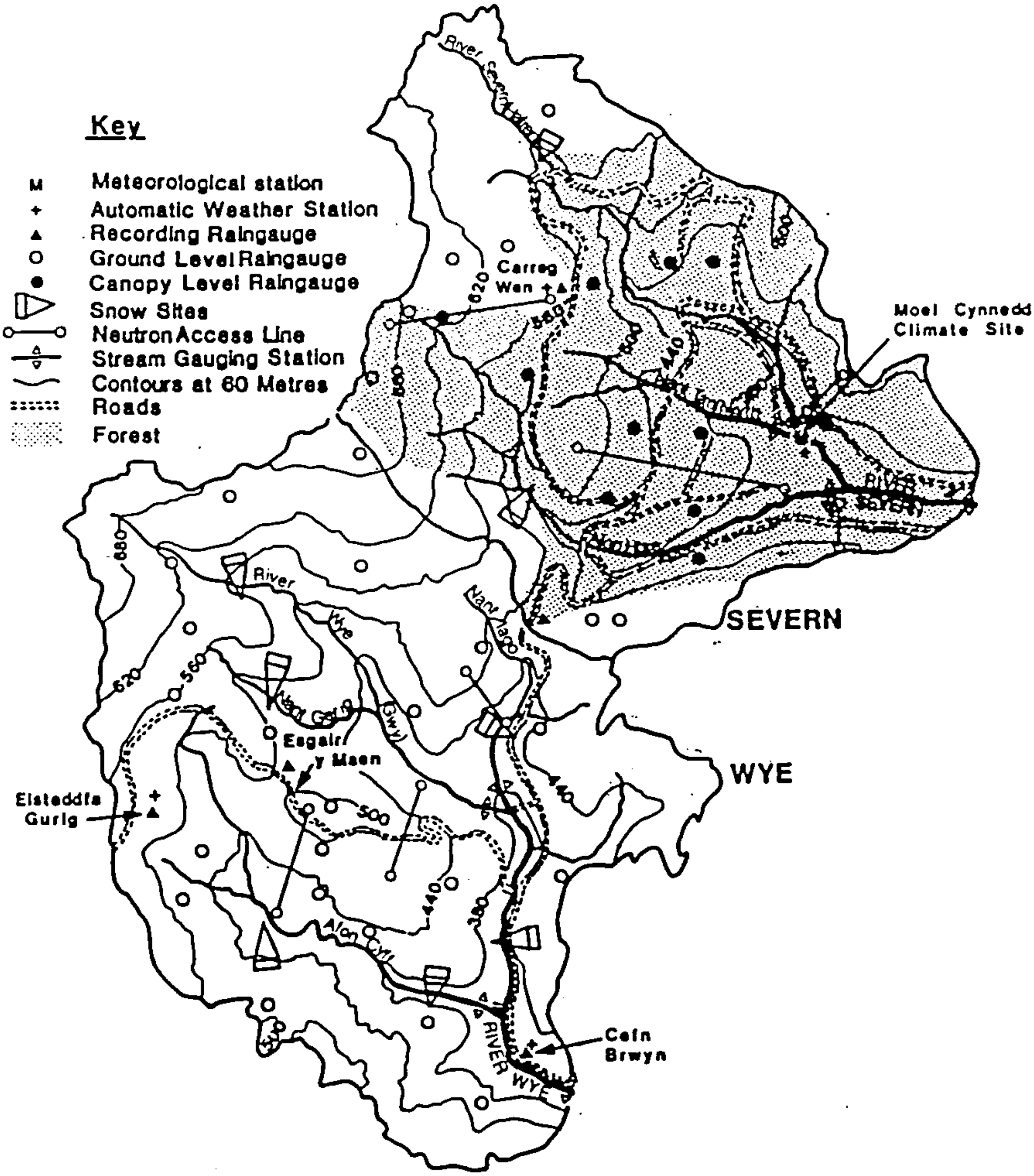


Figure 5 - The Plynlimon research catchments

approached through different disciplines. These studies are described extensively in the literature, such as by Dozier (1992) who considers their experimental design using groundbased measurements, aircraft and satellites. In more detailed examples, some of the background to the HAPEX-MOBILHY programme and some of the results it has produced are discussed by Choisnel et al (1987) and by André et al (1988).

4. Errors, uncertainties and accuracies

Unforunately for hydrologists, meteorologists and others involved, determining the water balance requires measurements to be made mostly in the natural environment under field conditions. The carefully controlled laboratory setting is only occasionally available to them. These field measurements have to be undertaken in hostile surroundings which are continually changing with time, as the natural variations in weather, climate and vegetation unfold and geomorphic processes proceed. But now human activities impose further modifications. To sample these changes, and particularly the extremes they contain, the measurements are best made over long-time periods. The presence of the sensor should not alter the variable being observed and the instruments should be located at sites which are properly representative of the area or basin being sampled. Table 6 indicates some of the problems of point precipitation measurement. For out-of-river variables, the representivity problem has been eased somewhat by the advent of weather radars and satellites, through the images they provide of the fields of several variables.

TABLE 6

Problems of assessing the water balance from point precipitation measurements

1.	Spatial coverage is often incomplete
2.	Temporal coverage is often incomplete
3.	There are at least 54 different types of standard gauge in use in 136 countries covering about 90% of the land area of the globe and in addition a large number of different types of rain recorders
4.	Errors of measurement have not been determined for each gauge type
5.	Installation of gauges and their sites may not meet the required practice
6.	Changes have occurred in gauge exposure
7.	Gauges have been moved
8.	New types of gauge have been introduced without comparisons with old versions
9.	Observer practice has altered
10.	Station histories are not documented

However, these images need measurements from strategically placed ground-based instruments for their calibration and interpretation. In other circumstances, for example in taking soil water content and water quality measurements, the representivity problem remains. Concentrations of constituents vary vertically and horizontally in a water body, and of course in time. Depth-integrated samples go some way to overcoming these difficulties and obviously the more samples taken, the more nearly the samples represent the whole (USGS 1993). These difficulties have often led to very dense networks being established initially, networks which are later reduced as hydrological patterns become established.

These problems and a number of others, such as those concerned with the storage and analysis of the data captured from the field, introduce errors into the measurements of the hydrological variables. Of course the error is strictly the difference between the result of a measurement and the true value of the quantity being measured (WMO 1994b). But for most hydrological variables, as for many other environmental variables, the true value is unknown and even a best approximation, such as might be obtained in the laboratory from a series of repeated measurements, is not usually available. This has led to a preference for the use of the term uncertainty--the interval about the measurement within which the true value of the quantity can be expected to be with a stated probability (WMO 1994b). Of course, hydrological measurements may contain all three types of errors or uncertainties, a spurious error may come from a raingauge being filled artificially, random errors may result from observer practice and a systematic error can come from the zero of a water level recorder not being set properly. But there are many other sources of error and Fig. 6 shows how these may arise in flow measurement with a weir or a flume. The overall uncertainty of measurement X_Q (Ackers et al 1978) depends on the standard of construction of the gauging structure, the correct application of the design specifications and a number of other factors and X_Q will also vary with discharge for a given structure. Herschy (1978 and 1985) discussed the errors of the various methods of flow measurement drawing on material published by the International Organization for Standardization (ISO) and also by WMO. He concluded that if ISO standards are followed, then flow measurements made at a single gauge are expected to have the following upper limits of uncertainties at the 95 % confidence limits:

Single determination of discharge	±7%
15 minute average of discharge	±5%
Daily mean, monthly mean or annual discharge	±5%

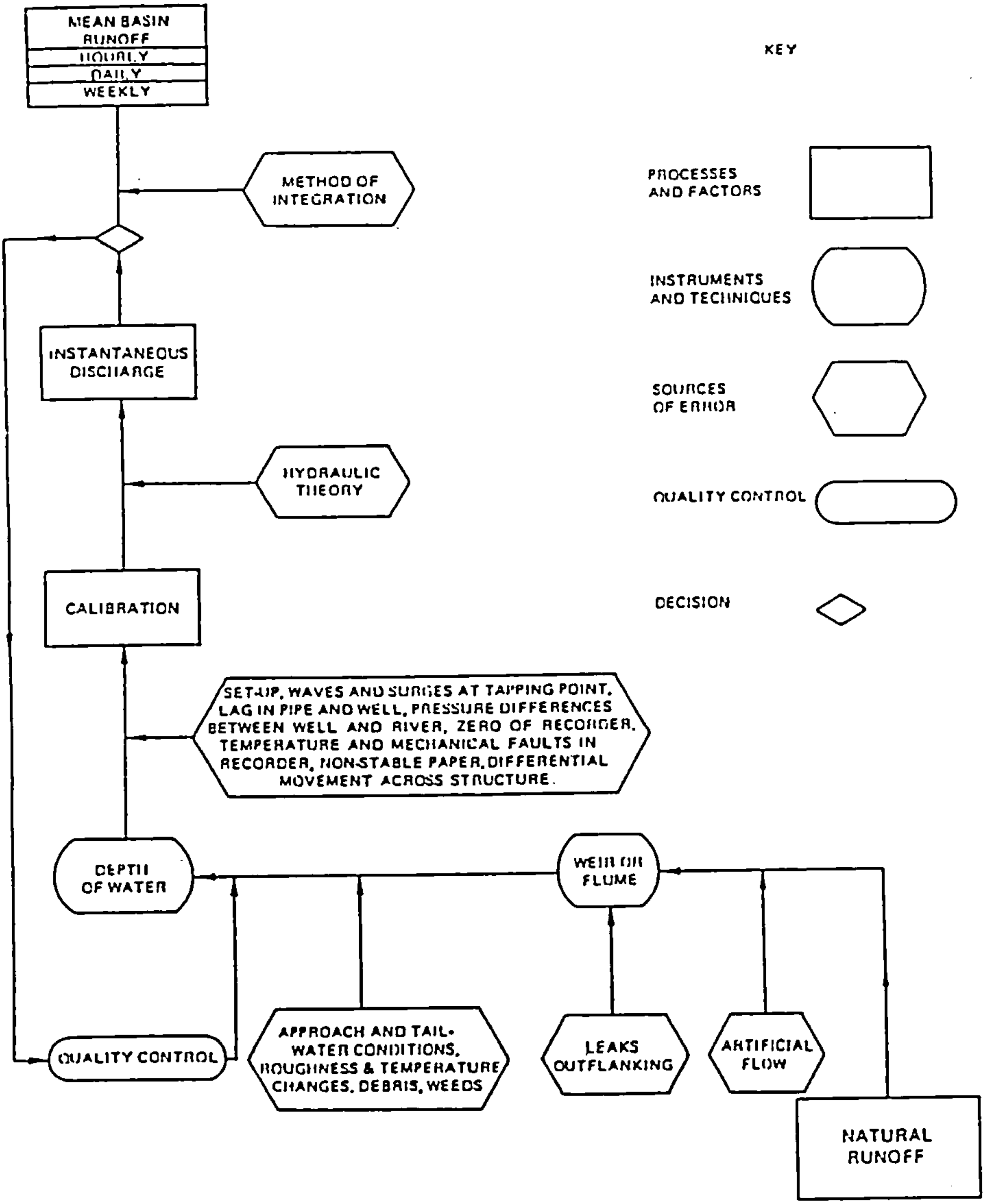


Figure 6 - Source of errors in the measurement of flow by a weir or flume

Of course, it would be most unwise to assume that river flows generally are measured to these limits: many countries are not able to meet ISO Standards (ISO 1983), nor the requirements of WMO (WMO 1988). Neither do the measurements of the other variables satisfy the WMO recommendations for accuracy (Table 7) that have been adopted by the WMO Commission for Hydrology. Unfortunately, these are accuracies to aim at rather than those being presently achieved: indeed the actual measurement errors must be somewhat larger than those stated. Take precipitation as an example: errors can range from 5 to 30% for rain and 10 to 80% for snow depending on gauge type, site, wind speed and other factors.

5. Reducing errors

For most of its history, WMO has been working in various ways to combat errors in instrument practice and in the errors which can occur at later stages in data management. One of the main vehicles for this work has been the series of instrument intercomparisons that cover the measurement of the main hydrological variables. For the first comparison of liquid precipitation measurements, one particular reference gauge was selected for the tests undertaken in the 1950s. When the faults of this gauge were recognized, a different reference was chosen for the second intercomparison made in the 1970s (Sevruk and Hamon 1984). For the third intercomparison, which is of solid precipitation measuring devices and is just concluding, a third interim reference was employed (Goodison et al 1994). These reference instruments were installed at a large number of different sites around the world for comparisons with the various national gauges.The results of the different intercomparisons show how national gauges performed against the reference. For the comparison of methods of evaporation measurement and estimation (WMO 1966) which was conducted in the 1960s, a large amount of data was collected from national tests and by means of a questionnaire. For Phase 1 of the intercomparison of water level recorders and current meters, tests were conducted nation by nation of their own instruments and several other instruments against the detailed test specifications which had been previously agreed (Starosolszky & Muszkalay 1978). For Phase 2 of this intercomparison, which was broadened to include data loggers, suspended sediment samplers and several categories of water level and velocity sensors, additionally the US Geological Survey made available a number of US P61 sediment samplers to participating countries for tests against national samplers (Wiebe 1994). Fig. 7 shows the results of the comparison of the US P61 and the equivalent German sampler. The other

TABLE 7

Recommended accuracies (uncertainty levels) expressed at the 95 per cent confidence interval for different variables*

Variable	Recommended accuracy
Precipitation (amount and form)	3-7%
Rainfall intensity	1 mm/h
Snow depth (point)	1 cm below 20 cm or 10% above 20 cm
Water content of snow	2.5-10%
Evaporation (point)	2-5%, 0.5 mm
Wind speed	0.5 m/s
Water level	10-20 mm
Wave height	10%
Water depth	0.1 m, 2%
Width of water surface	0.5%
Velocity of flow	2.5%
Discharge	5%
Suspended sediment concentration	10%
Suspended sediment transport	10%
Bed-load transport	25%
Water temperature	0.1-0.5°
Dissolved oxygen (water temperature is more than 10°C)	3%
Turbidity	5-10%
Colour	5%
pH	0.05-0.1 pH unit
Electrical conductivity	5%
Ice thickness	1.2 cm, 5%
Ice coverage	5% for ≥ 20 kg/m³
Soil moisture	1 kg/m³ ≥ 20 kg/m³

* NOTE: When a range of accuracy levels is recommended, the lower value is applicable to measurements under relatively good conditions and the higher value is applicable to measurements in a difficult situation.

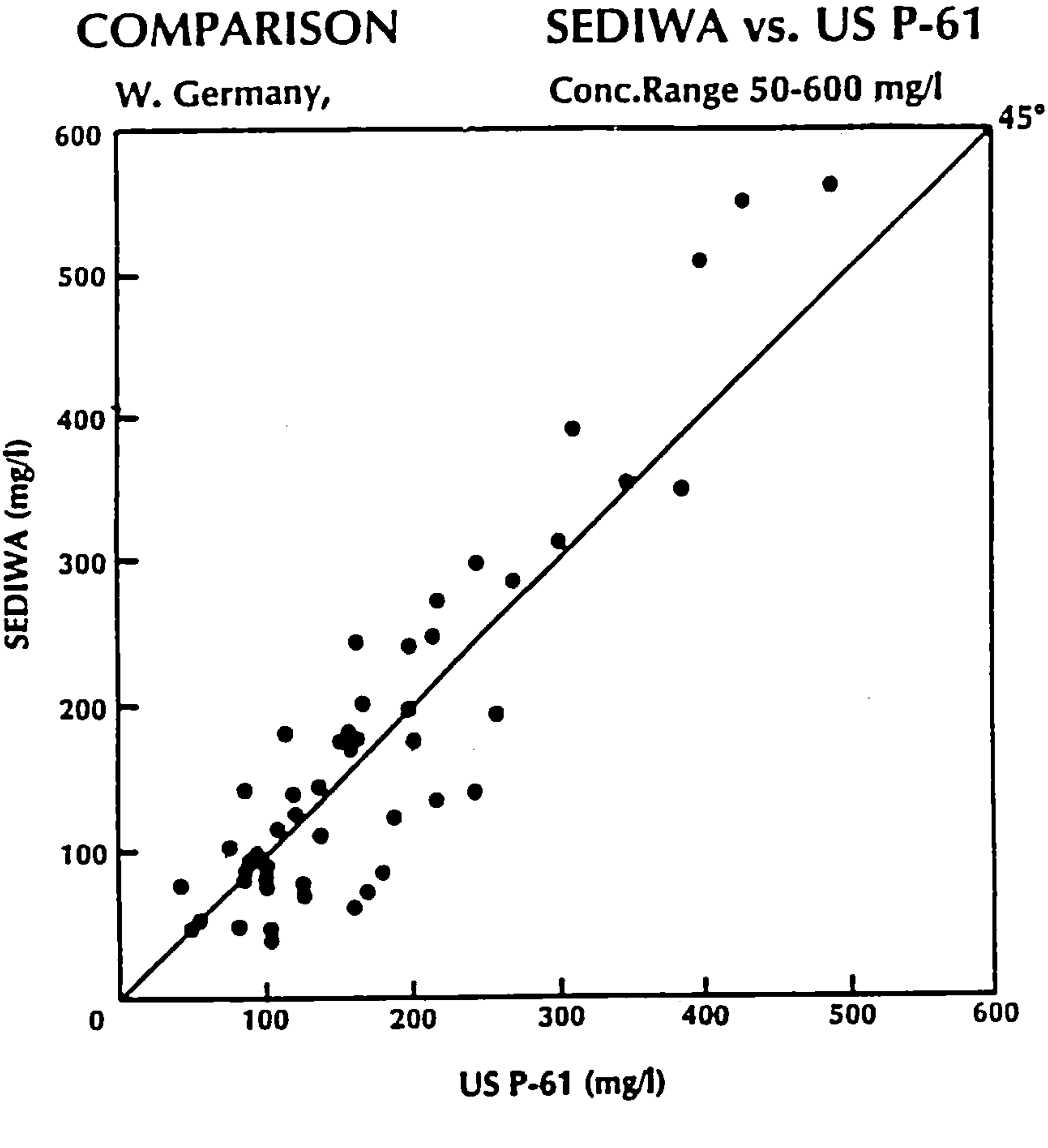


Figure 7 - Intercomparison of suspended sediment samplers

results of these intercomparisons show how different instruments performed against the test specifications.

Unfortunately, many different interests mitigate against the formulation of the results of these intercomparisons in the manner of a consumer report on the 'best buy'. So no clear preferences for devices of particular types are stated in the conclusions. However, for the discerning reader, it is usually obvious which of the instruments performed most closely to the specifications of the tests but it was not found possible to include in the results a statement of the errors of measurement for the variables concerned. However, the tests themselves, through the transfer of technology they promote, help to raise the standard of performance of the hydrological services involved, which in turn assists in error reduction in the long term. The same applies to the transfer of technology facilitated by HOMS (the Hydrological Operational Multipurpose System), the WMO technology transfer system which has achieved more than 3000 exchanges of components between hydrologists in developed and developing countries. But neither as a result of these comparisons, nor in HOMS does WMO, nor any other organization, advocate the use of a particular type of instrument, so that observations might be made in a uniform way world-wide. If this were the case, and all the subsequent data management procedures were also harmonized, then there would be few difficulties in ascribing errors to the estimates of water balances. This type of observational uniformity is what the GEMS Water Quality Programme (WHO 1991) aims to achieve. It is what has been achieved by certain national programmes in water quality, for example; the United States National Water Information System (USGS1993) and the United Kingdom's Harmonized Monitoring System (Brown et al 1982).

6. Total quality management

Perhaps the time is ripe to adopt similar initiatives globally and to embrace the concept of total quality management which is embodied (ISO 1993) in the ISO 9000 series of standards (ISO 1987)(ISO 1973). These offer an integrated global system for optimizing the quality effectiveness of a company or organization. By adhering to defined standards, an organization ensures that its product is "fit for the purpose for which it is intended", a hydrometeorological data set being one such product. This is the approach that has been adopted by New Zealand's National Institute for Water and Atmosphere (NIWA) (Mosley & McMillan 1994) for its environmental data activities, including hydrological data. The reasons were:

1. to eliminate the costs of poor quality control;

2. to ensure that the product meets the user's needs that the data are confidently usable;

3. to demonstrate to users that the organisation is meeting and anticipating their needs.

The approach incorporates adherence to manuals and procedures (Table 8) and includes certification by an independent body accredited to carry out the inspection work. NIWA implemented new procedures and upgraded instruments and equipment to meet the necessary requirements the result was certification which was obtained in April 1992. The end products are data sets which achieve the standards of quality claimed for them.

7. Concluding remarks

Are hydrological data for New Zealand the only ones globally which possess the attributes of a known quality standard? Can the ISO 9000 ideas be implemented generally? Perhaps the establishment of WHYCOS is the key to the propagation of quality assurance for hydrological data world-wide, with the operation of the stations and the procedures being designed to meet the ISO and WMO requirements. With this approach, it will be possible at some future date to weigh up the water balance basin-wide and globally. Then the amounts of water involved and the loads they carry can be prescribed fully. Now despite the coming water crisis we can only make educated guesses.

Acknowledgements

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TABLE 8
Quality system reporting procedures

Report	By	For	When	Frequency	Purpose
1. Quarterly report on the state of archived data	Water Resources Databank (WRD)	General Manager, ED	First week in following quarter	Quarterly	Summarize data quality; indentify variances
2. Annual report on archive data	WRD	General Manager, ED	By 30 September of following year	Annual	Summarize data quality; identify variances
3. Annual station data review	ED units Manager, WRD	Manager, QA/SD	By 30 June of following year	Annual	Illustrate data quality, identify variances
4. Annual station inspection	ED Units	Manager, QA/SD	Continuous	Biennial	Indicate compliance to field practice standards, identify discrepancies and remedies
5. Biennial station inspection	QA/SD Unit	QA/SD Unit	Continuous	Biennial	Ensure adherence to field and office practice standards, identify discrepancies and remedies
6. Assessment of field practice	ED Units	Manager, QA/SD	Continuous	Annual	Ensure adherence to field and office practices, identify training needs etc.
7. Random discharge measurements	QA/SD Unit	QA/SD Unit	Intermittent	< Annually	Ensure data quality; identify variances
8. Annual QA system review	QA/SD Unit, General Manager ED, and Manager, WRD	Chief Executive Officer, NIWAR	By 30 October of following year	Annual	Indicate effectiveness of QA system, and ensure attainment of all standards of data quality and practices
9. Between unit discharge measurement	Water Resources Survey Units	QA/SD Unit	Intermittent	Tri-annual	Demonstrate adherence to field practices and standards, meter calibrations and accuracies
10. ED Support Centre (EDSC) fault analysis	EDSC	Manager, QA/SD	By 30 June of following year	Annual	Identify causes of instrumentation failures, including operator problems

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FIFTY YEARS OF POTENTIAL EVAPORATION

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Abstract

The concept of "potential" evaporation (PE), now 50 years old, was central to equations derived by Thornthwaite, Penman, Priestley and Taylor and others. Penman's equation has proved the most versatile. When modified by introducing a canopy resistance to estimate PE for well-watered vegetation, it performed well in recent tests against world-wide measurements with lysimeters. Bouchet's complementary relation has been used to estimate sub-potential rates of evaporation from PE. A new scheme is described in which rates of evaporation depend on physical feedback in a Convective Boundary Layer and physiological feedback within plants.

History

Potential Evaporation is an appropriate topic for this meeting because the subject's jubilee falls this year. In a paper published in 1944 and now little known, the American climatologist C.W.Thornthwaite used the term "Potential Evapotranspiration" defined as "the loss of water from a moist soil tract completely covered with vegetation ... and large enough for oasis effects to be negligible" (Thornthwaite, 1944, as quoted by Thornthwaite and Hare, 1965). Four years later, the concept of potential evaporation was expounded and explored in two major papers, still frequently quoted: " An approach towards a rational classification of climate" by Thornthwaite (1948) and "Natural evaporation from open water, bare soil and grass" by Penman (1948). Because Penman's treatment was based on quantitative physical principles, it has stood the test of time better and few, if any, papers in agricultural meteorology and hydrology have been cited so frequently throughout the second half of this century.

Although Penman (1948) did not cite Thornthwaite (1944) it is most unlikely that he was unaware of the paper. Subsequently, he acknowledged that Thornthwaite had introduced temperature "as a parameter that would integrate the balance of radiation exchanges, air movement and other meteorological factors that might affect evaporation (Penman, 1963)". He was less complimentary about the term "evapotranspiration" which his colleagues were never allowed to use: "...it is presumably a useful term but it is rather ugly and it hardly seems necessary, as there are few situations in which the use of "evaporation" or "transpiration" is not entirely adequate" (Penman, 1963). The term persists however, particularly across the Atlantic where, in 1985, I heard a senior climatologist talking about "evapotranspiration from the oceans!"

What did Thornthwaite think of Penman's work? Almost certainly he admired it - but not publicly. At a meeting in Aspendale, Australia in 1956, he announced, tongue at least partly in cheek, that he found Penman's formula "too empirical". "How then does your own formula work so well, Dr.Thornthwaite?" "I don't really know; it's just magic"!

Standard formulae for potential evaporation

THORNTHWAITE

Thornthwaite's (1948) formula was based on rainfall and runoff measurements for river basins in the USA. It is economical in climatic parameters (monthly mean temperature T in ^{o}C only); but it is prodigal in numerical constants and in the precision to which they are expressed. In its original form, potential evaporation for a month (in mm) was given as

$$PE = 1.6(10T/I)^{a} \tag{1}$$

where $$a = (0.675I^{3} - 77.1I^{2} + 17920I + 492390)\times10^{-6} \tag{1a}$$

and $$I = \sum (T/5)^{1.514} \tag{1b}$$

where the summation in eqn 1b is over 12 months.

No one has criticised his formula more trenchantly than Thornthwaite himself! In the Appendix to his 1948 paper he wrote: "This mathematical development is far from satisfactory...it is completely lacking in mathematical elegance. It is very complicated and without nomograms and tables as computing aids would be quite unworkable. The chief obstacle at present to the development of a rational equation is the lack of understanding of why potential evapotranspiration corresponding to a given temperature is not the same everywhere." Later, Thornthwaite (1954) warned that estimates of PE from his formula diverged from measurements in regions where the climate was markedly different from those used in its derivation.

With the passage of time, these caveats and misgivings seem to have been forgotten and his equation is still being used, for example, in attempts to estimate evaporation from satellite radiometry, justified by the fact that air temperature is more widely reported than any other climatological quantity apart from rainfall.

PENMAN

Penman's formula for PE is more elegant than Thornthwaite's and has a sound physical base. In its most abbreviated form (Penman, 1963), the latent heat of evaporation, λE, from a uniform surface over any prescribed period is

$$\lambda E = \frac{(\Delta/\gamma)H + E_a}{\Delta/\gamma + x} \quad (2)$$

where H (Penman's original notation) is the net heat available from radiation and from any change of heat stored in the system (if any). In contemporary notation, the quantity E_a is $CD/(\gamma r_H)$ where r_H is a wind-dependent resistance for the transfer of sensible heat between the surface and a reference height z_r above it, C is the volumetric specific heat of air at constant pressure, and D is the mean saturation deficit of air at height z_r. The familiar parameters Δ and γ are temperature-dependent properties of moist air in consistent units.

The quantity x is the ratio of the total transfer resistance for vapour, r_V, to r_H. To find the potential rate of evaporation from an open water surface (PE(w)), x = 1 is an acceptable approximation (although r_H and r_V are not precisely equal). For regional or global use, the main disadvantage of Penman's formula is the number of climatic measurements needed -temperature, humidity, wind and net radiation which often has to be estimated from hours of sunshine or cloudiness because reliable climatological measurements of this quantity are still rare.

Penman defined potential transpiration for vegetation (PE(v)) as the rate of evaporation from an extensive short, green crop, fully covering the ground and never short of water. He determined PE(v) as a fraction of PE(w) which changed during the year (from about 0.6 to 0.8 in Western Europe); and he estimated x from daylength and stomatal geometry (Penman, 1952).

One of the first large-scale tests of the Penman formula revealed excellent agreement between the annual excess of rainfall over runoff for river basins in the Britain Isles and estimated annual PE (Penman, 1950). The Meteorological Office was not convinced by this evidence, however, and Pasquill (1950) claimed that "application of the method to the broad estimation of annual land evaporation is no more reliable than a simple and more direct approach in which... (the latent heat of) evaporation is equated to the net income of radiant energy". This criticism did not worry Penman and may, in fact have stimulated him to demonstrate the validity of his formula for a wide range of catchments and climates (Penman, 1963).

Later, it was shown that when a uniform stand of vegetation was treated as a "big leaf", the total resistance to vapour transfer from vegetation could be expressed as the sum of two components, one depending on windspeed and the aerodynamic properties of the surface (r_a) and the other on "canopy" or "surface resistance" (r_s), a quantity depending on the distribution and size of stomata (Monteith, 1965). Because of the difficulty of estimating r_s *a priori*, the Penman-Monteith (PM) equation has usually been employed diagnostically to calculate canopy resistance when the transpiration rate and other variables are known. As a guide, to potential rates of transpiration , the smallest recorded values of r_s are around 50 sm^{-1} for both field crops and coniferous forest (Kelliher, et al., 1993). Corresponding characteristic values of aerodynamic resistance are around 20-40 sm^{-1} for crops and 5 sm^{-1} for forests. It follows from eqn 2 that estimates of PE for forests are relatively sensitive to vapour pressure deficit D and to maximum stomatal conductance whereas the PE of short vegetation is more responsive to radiant energy and less responsive to maximum conductance. Jarvis and McNaughton (1986) examined these differences in terms of the extent to which different types of vegetation are "coupled" to the atmosphere.

In the 1970s, the validity of conventional PE calculations for forests was called into question by workers at the Institute of Hydrology, Wallingford, who had estimated faster rates from the water balance of forested catchments. The discrepancy led to a major re-examination of Penman's formula by Thom and Oliver (1977). They showed that the wind function or "aerodynamic conductance" used in the standard Penman formula (derived from measurements with a small, non-standard open-water tank), underestimated conductance for arable crops and grossly underestimated it for forests. They therefore proposed that the PE for areas mainly under arable cropping should be estimated by setting $x = 2.4$ (corresponding to $r_s/r_a = 1.4$) and by dividing the value of r_a as calculated by Penman by a factor of 2.5. These two changes are partly self-cancelling. For forests (see above), r_s/r_a was of the order of 10. The net consequence of these differences is to make annual potential and actual evaporation from forested areas in the UK about 50 to 100% larger than estimates from the original Penman formula.

Several others workers have attempted to improve the reliability of the Penman formula by altering the wind function. One of these variations, suggested in an FAO Report by Doorenbos and Pruitt (1977), is often referred to as FAO-24. Another proposed by Wright and Jensen (1972) based on lysimeter measurements at Kimberly, Idaho, is known as the Kimberly Penman equation. Tests of these equations are reported later.

RADIATION-BASED FORMULAE

The method developed by Budyko (1956) for estimating potential evaporation from wet land surfaces was essentially the same as Penman's but annual net radiation R_n was estimated at air temperature ("isothermal" net radiation) and the primary equations were solved by a process of iteration to avoid the approximation of evaluating Δ at air temperature. From an extensive set of data, Budyko found that R_n could be estimated from the sum of daily mean temperature above a base of $10^{o}C$ and, following Pasquill (1950), estimated annual evaporation as R_n/λ.

Makkink (1957) showed that PE over the Netherlands could be estimated as a linear function of *solar* radiation and Doorenbos and Pruitt (1977) tabulated the dominant coefficient of the associated regression as a function of daytime windspeed and mean relative humidity. This method is attractive because climatological statistics for net radiation are very rare.

PRIESTLEY-TAYLOR AND SLATYER-McILROY

In the final paragraph of his monograph "Turbulent Transfer in the Lower Atmosphere", Priestley (1959) briefly discussed the special case of heat transfer from a fully wet surface into a saturated atmosphere. The ratio of latent to sensible heat is then given by Δ/γ and if the sum of sensible and latent heat is H it can be shown that the latent heat equivalent of the potential evaporation rate is

$$\lambda E = \frac{\Delta H}{\Delta + \gamma}. \quad (3)$$

When the surface is wet but the air above it is not, λE is larger than this value.

To extend the applicability of this equation, Slatyer and McIlroy (1961) specified the effective wetness of a surface by the wet-bulb depression δ_0 of air in contact with it. Then if δ is the corresponding depression at a reference height for which the corresponding aerodynamic resistance is r, the latent heat equivalent of the **actual** evaporation rate given by the Slatyer-McIlroy (SM) equation is

$$\lambda E = \frac{\Delta H}{\Delta + \gamma} + (C/r_H)(\delta - \delta_0) \quad (4)$$

A **potential** rate of evaporation was obtained by simply setting $\delta_0 = 0$.

Later, Priestley and Taylor (1972) concluded from an exhaustive review of evaporation measurements for open water and for well-watered vegetation that the second term in eqn 4 was about 20 to 30% larger than the first term. They therefore suggested that the Penman and Slatyer-McIlroy formulas for PE could both be replaced by

$$\lambda E = \frac{\alpha \Delta H}{\Delta + \gamma} \quad (5)$$

where α, the so-called Priestley-Taylor coefficient, had a recommended value of 1.26.

When many subsequent workers working in temperate or cold climates confirmed values of α in the range 1.2 to 1.3, it appeared that the Priestley-Taylor (PT) equation might supersede Penman's equation as a more convenient and equally reliable method of estimating PE. However, from lysimeter measurements on potato and lucerne crops in Wisconsin, Jury and Tanner (1975) found that eqn 5 systematically underestimated PE during dry spells and proposed an additional term based on the ratio of the actual saturation deficit to "the long-term mean saturation deficit for a period when advection is low". However, adjusting the PT equation in this empirical way is tantamount to developing an inferior version of the Penman equation. Gunston and Batchelor (1983) drew attention to the fact that in semi-arid tropical climates, the ratio of PE calculated by the (unadjusted) PT equation to Penman's equation ranged from 1 to 3 depending on season.

The PT equation also grossly underestimates PE(v) from very rough surfaces such as forests with wet foliage, as shown, for example, by Shuttleworth and Calder (1979).

COMPARISON OF FORMULAE

Four major methods of estimating PE for a well-watered surface have been briefly reviewed: in historical order, Thornthwaite's (eqn.1), Penman's (eqn 2 with $x = 1$), Slatyer and McIlroy's (eqn 4 with $\delta_0 = 0$), and Priestley and Taylor's (eqn 5 with $\alpha \approx 1.26$). Using measurements reported by McIlroy and Angus (1964), Sellers (1965) compiled a table (loc.cit.p.173) in which monthly estimates of transpiration by the first three methods were compared with measurements from a lysimeter at Aspendale (Victoria, Australia) on which short grass was grown. Estimates from the Priestley-Taylor equation were obtained from tabulated values of Rn/λ and of mean air temperature.

Fig.1 presents this data graphically. Both the SM and Penman equation performed well but the former tended to overestimate transpiration in the wet winter months whereas the latter underestimated in dry summer months. The PT equation underestimated PE throughout the year and the Thornthwaite equation grossly underestimated.

McIlroy and Angus (1964) discussed the significance of local and regional advection of sensible heat in determining evaporation rates at Aspendale. It appears that both processes were accounted for by the humidity term in the Penman and SM equations but not in the (unadjusted) PT equation in which this term is absent. When regional advection is minimal, as in the centre of the Amazonian forest, both Penman and PT equations provided estimates of evaporation that were very close to measured rates throughout the year (Shuttleworth, 1988).

In a more recent and very comprehensive comparison of Penman-type methods for estimating PE(v), Allen et al (1989) used measurements of evaporation from grass and lucerne grown in

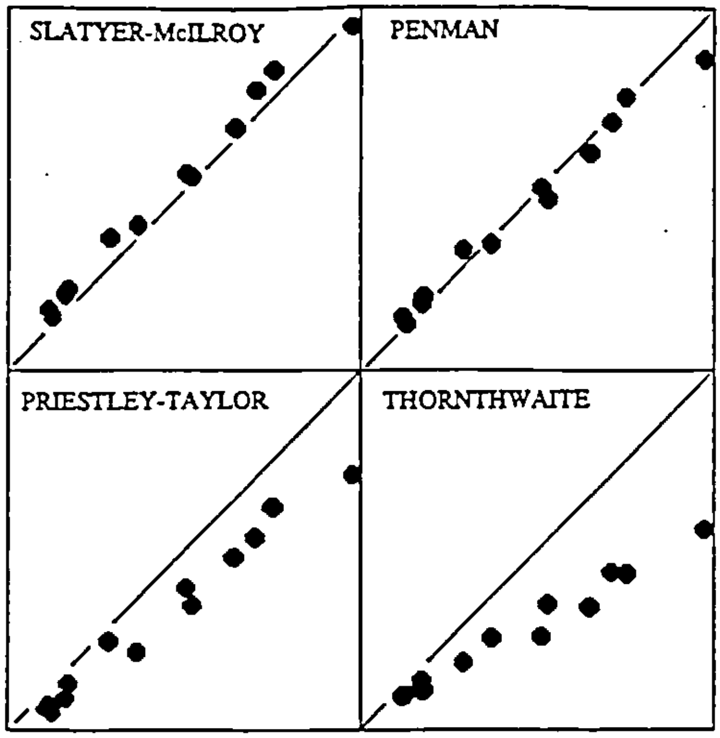

Fig.1 Comparison of estimated and measured rates of monthly evaporation from well-watered short grass at Aspendale, Australia, after McIlroy and Angus (1964) and Sellers (1965). In each box, the vertical (estimate) and horizontal (measurement) scales are 0 to 8 cm per month.

well-managed lysimeters at 11 world-wide sites covering a range of climates. Table 1 of the Allen et al study contains a sub-set of the output from this extensive study in the form of standard deviations of monthly PE estimates compared with measurements. For all locations and months, the original Penman, PM, and Kimberly Penman equations yielded deviations in the range 0.4 to 0.6 mm d^{-1}. The FAO-24 version performed worst with an SD of 1.17mm d^{-1}. (From inspection, the mean PE appears to be about 4 mm d^{-1} so these absolute deviations are equivalent to about 10 to 15% for the three best formulae and 30% for the worst).

The Penman equation tended to overestimate slightly at small values of PE and vice versa so that the slope of the regression (estimated versus measured PE) was m = 1.01 ($r^2 = 0.94$). The Kimberly-Penman equation also overestimated at small PE (m = 1.08, $r^2 = 0.92$) and FAO-24 overestimated over the whole range of PE (m = 1.22, $r^2 = 0.92$). For the PM equation, m = 1.00 ($r^2 = 0.96$).

The main difference between the four formulae lies in the wind function adopted. In this respect, the success of the original Penman equation is remarkable, considering the primitive nature of the tank measurements from which Penman derived his wind function. The success of the PM equation lies partly in the dependence of the wind function on well-established aerodynamic parameters and partly on the addition of a physiological term. The equation is therefore the most reliable of the four but needs more input. Despite this disadvantage, a panel of experts meeting in Rome in 1990 recommended that the PM equation should replace FAO-24 in a future edition of the FAO handbook for Crop Water Requirements (FAO,1991).

Interaction between vegetation and the atmosphere

COMPLEMENTARITY

The net amount of radiant energy absorbed by water, soil or vegetation is dominated by the amount of solar energy received and partitioning of that energy between sensible and latent heat depends on the availability of water. During a spell of rainless weather, a progressive decline in the transfer of latent heat from surface to atmosphere is therefore matched by an increase in the transfer of sensible heat. Provided there is no major synoptic change, the atmosphere becomes warmer and drier so that saturation deficit increases and **potential** rates of transpiration increase as **actual** rates decrease. The scale of this process is often magnified by an increase in the entrainment of relatively warm, dry air across the inversion that caps the Convective Boundary Layer.

Bouchet (1963) put forward the hypothesis (often referred to as the complementarity relation or CR) that an increase in **potential** rate of evaporation E_p should be **equal** to the decrease in actual rate E_a. Then if E_{po} is the potential (= actual) rate of evaporation when the supply of water is unlimited, it follows that

$$E_a + E_p = 2E_{po} \qquad (6)$$

so that E_a can be estimated as $2E_{po}$ - E_p. In further analysis, Bouchet suggested that $2E_{po}$ could be taken as the net absorbed solar radiation at a site. No quantitative theoretical basis has yet been found for these essentially intuitive relations.

The type of complementarity explored by Bouchet was conspicuous during the long UK drought of 1976 as shown by Fig 2 in which tank evaporation appears as a surrogate for potential evaporation and the evaporation from the Meteorological Office grass lysimeter at Cardington as a surrogate for the average actual evaporation from Bedfordshire!.

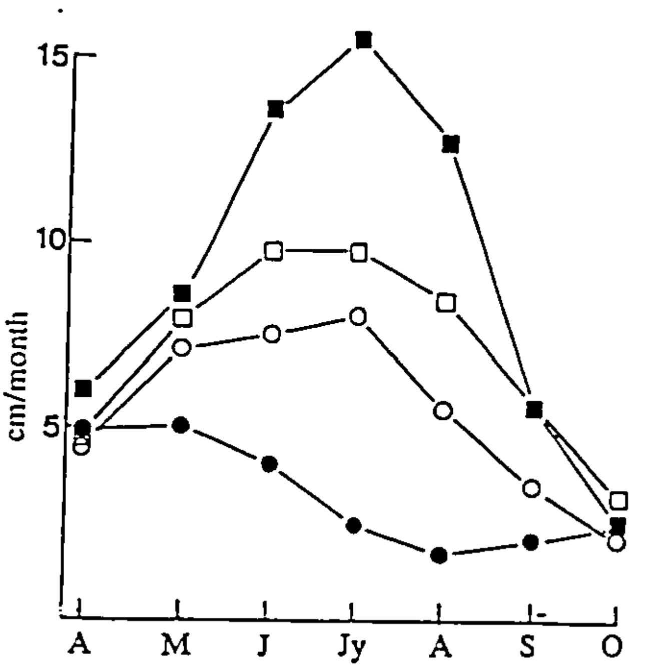

Fig.2 Monthly rates of evaporation from a grass lysimeter and open water tank at Cardington, Bedfordshire in a very dry year (1976) and averaged for years of normal rainfall (1970-75) (after Richards, 1976).

	water	grass
1976	■	●
1970-75	□	⊙

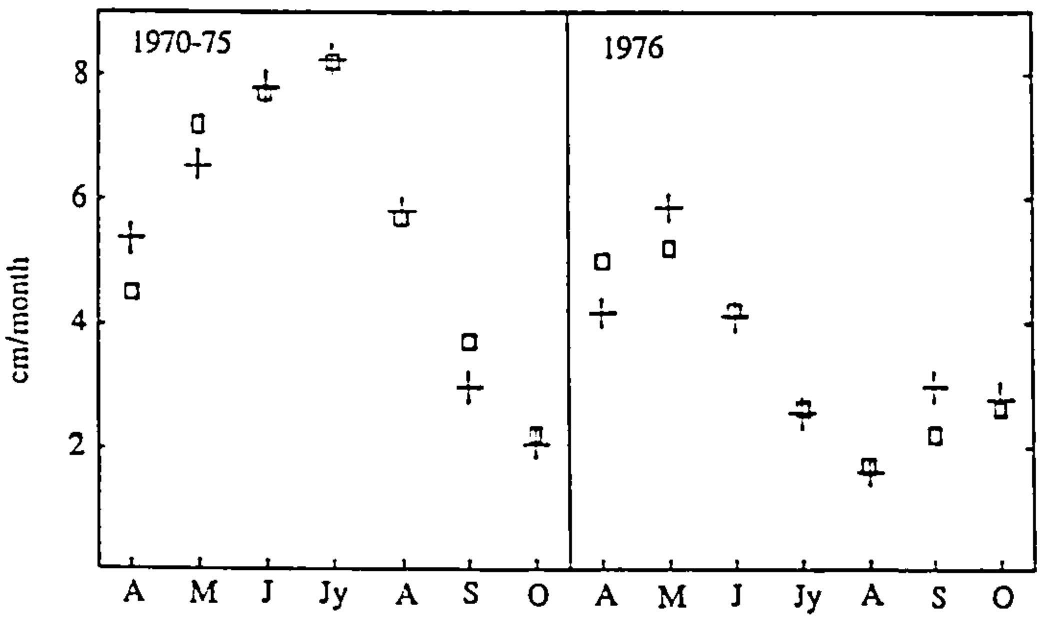

Fig.3 Monthly rates of actual evaporation at Cardington, Bedfordshire as measured with a grass lysimeter (□) and estimated from potential rates (+) as described in text.

In the years 1970-75 when monthly rainfall rarely departed far from the average, actual and potential rates of evaporation changed with time in the same direction (because both were driven by the supply of radiant energy) (Fig.2). In 1976, however, they changed in opposite directions because of the advection of sensible heat from the dry area surrounding the tank. However, consistently with the CR, the sum of actual and potential rates was almost independent of rainfall distribution. Taking monthly averages of this sum for both types of season ("normal" and exceptionally dry) and subtracting measured open water evaporation provides the estimates of actual evaporation shown in Fig.3. Bearing in mind the surrogate nature of the measurements on which the estimates are based, the agreement between actual (lysimeter) and predicted evaporation is surprisingly good in both types of season and discrepancies reveal no bias. Using tank measurements of PE(w) to estimate actual evaporation in this way may merit further investigation. Brutsaert and Stricker (1979) tested the CR for a catchment in eastern Holland in 1976 by using the Penman equation to find E_p and the Priestley-Taylor equation to find E_{po}. For this set of observations, agreement between predicted and measured rates of evaporation was "generally good" (Brutsaert, 1982).

Morton (1983) has made much wider use of the CR to analyse the water balance of major river basins throughout the world obtaining extremely good agreement between estimated evaporation and the difference between rainfall and runoff. In an early study of 20 catchments in Ireland and North America, ranging in area from 97 to 14370 km^2, all estimates of annual evaporation were within about ± 10% of measurements (Morton, 1971). However, making the conservative assumption that errors in rainfall and runoff were of the order of 5 to 10%, error expected in the difference between these quantities (i.e. in "actual" evaporation) would be of the order of 10 to 20%. Errors of similar magnitude are associated with estimates of evaporation based on (a) climatological information at selected sites that are unlikely to be perfectly representative of the whole catchment; and (b) on empirical relations, e.g. between the input of radiant energy and cloudiness. The close agreement between prediction and measurement in this and similar studies may therefore be a consequence of using empirical functions (e.g. for advection) to calibrate the model as described by Morton (1983).

THE CONVECTIVE BOUNDARY LAYER

Within the past decade, several workers have developed models of the convective boundary layer (CBL) that offer new ways to explore the relation between potential and actual rates of evaporation from the surface. These models describe budgets for sensible and latent heat within a dry, cloud free-layer of the order of 1 to 3 km deep during the day and capped by an inversion above which potential temperature increases with height and the specific humidity decreases. Turbulence at the inversion, generated by buoyancy when there is an input of sensible heat from the surface, entrains warm dry air. Given the state of the atmosphere at surface, and the input of radiation, it is possible to estimate how temperature, humidity and the partitioning of radiant energy should change with time.

By combining a simplified CBL model with a set of surface and boundary layer measurements made in the Netherlands by Driedonks (1981), McNaughton and Spriggs (1989) were able to estimate notional values for the Priestley-Taylor coefficient (α) as a function of surface resistance r_s. (The coefficient was averaged over the period from dawn to mid-afternoon when boundary-layer depth was increasing and the resistance was assumed independent of time). The dependence of α on r_s changed from day to day according changes in gradients of temperature and mixing ratio at the top of the CBL. Fig.4(a) shows the relation for two days characterised by extreme values of α evaluated at zero resistance, i.e., α_0 = 1.3 or 1.6. The sigmoid nature

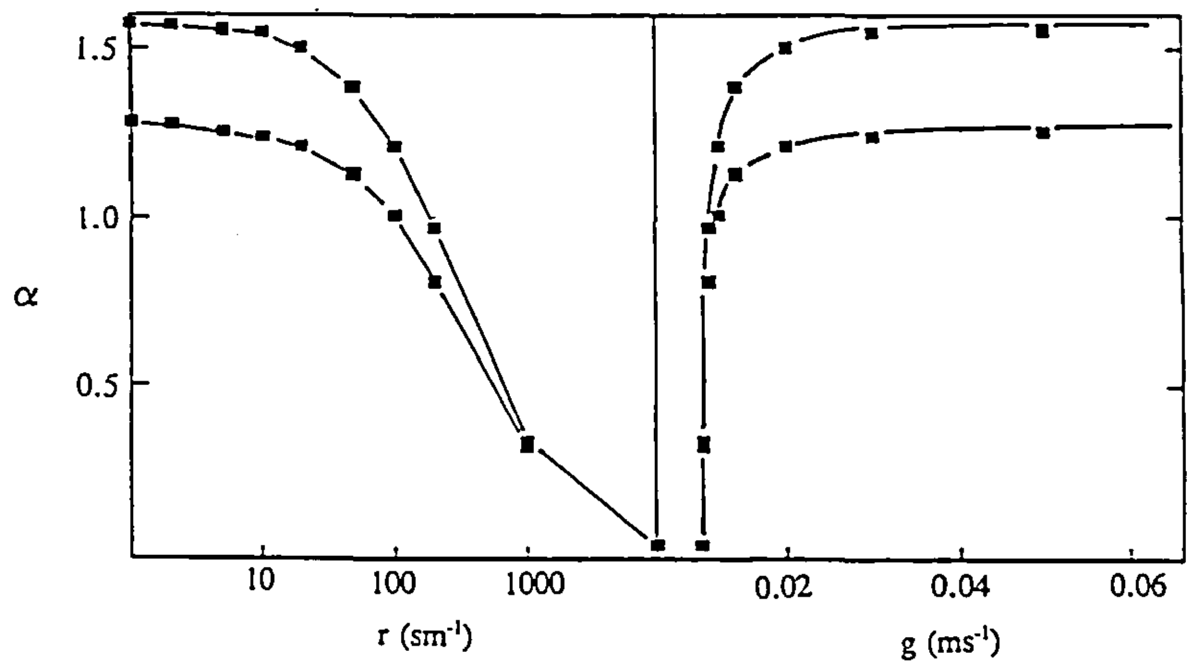

Fig.4 Dependence of Priestley-Taylor coefficient α on surface resistance r as estimated by McNaughton and Spriggs (1989) from Convective Boundary Layer measurements at Cabauw, Netherlands (Driedonks, 1981). (a): as plotted by McNaughton and Spriggs with log(r) as abscissa. (b): with g = 1/r as abscissa.

of the relation as plotted by McNaughton and Spriggs is a consequence of their choice of $\log(r_s)$ as the independent variable. A simpler and more informative relation can be obtained by plotting α against the surface *conductance* $1/r_s$ (Fig.4(b)) to obtain points that fit the hyperbolic relation

$$\alpha = (1/\alpha_0 + br_s)^{-1} \qquad (7)$$

(Monteith, 1994a).

From eqn 2, it can be shown that the parameter $b = d(1/\alpha)/dr_s$ depends both on the value of r_H and on the rate at which saturation deficit increases with r_s (dD/dr_s). As evaluated from the Cabauw data set, appears to be a very weak function of r_s. In effect, eqn (7) uses PE(w), which is proportional to $\alpha_0 H$, to determine both potential rates of transpiration (PE(v)) by setting r_s at some arbitary minimum value such as 50 sm^{-1}; and to determine actual rates of transpiration when r_s exceeds that value.

PHYSIOLOGICAL RESPONSE

In the analysis of CBL behaviour described by McNaughton and Spriggs (1989), surface resistance was prescribed and assumed constant with time. For vegetation, however, the resistance depends mainly on the extent to which stomata are open and therefore on factors such as solar irradiance, temperature, and the balance between water supply and demand. In most contemporary work, "supply" is assumed to be a function soil water content within the root zone and "demand" is a function of saturation deficit. A convenient supply equation consistent with a wealth of laboratory evidence (Monteith, 1994b) is

$$g = g_{max}(1 - E/E_{max}) \quad (8)$$

$$= g_{max}(1 - \alpha/\alpha_{max}) \quad (8a)$$

where g_{max} is a maximum stomatal conductance obtained when $E = 0$ (i.e. when there is no demand for water); E_{max} is a notional maximum rate of water supply (notional because when $E = E_{max}$, the equation implies that stomata must be closed); and

$$\alpha_{max} = \{(\Delta + \gamma)/\Delta\}\lambda E_{max}/H. \quad (8b)$$

Equation 8 has not yet been rigorously tested outside the laboratory but it is consistent with much field evidence that surface conductance decreases with saturation deficit (other factors remaining constant) and, in particular, with the evidence that surface resistance is a linear function of saturation vapour pressure deficit as suggested by Lohammer et al (1980) and others. Assuming that eqn 8 is valid in the field, solving eqns 7 and 8 simultaneously then gives surface conductance g and the Priestley-Taylor coefficient as functions of α_o, α_{max}, g_{max} and b. Eqn 2 can then be rearranged to find saturation deficit D as a function of these parameters. This scheme assumes that temperature (and therefore Δ) is independent of r_s which is unrealistic and iteration would be needed to obtain a more exact solution with temperature allowed to change.

BALANCING SUPPLY AND DEMAND

The nature of accommodation between the atmosphere and vegetation can now be explored in terms of water supply and demand. In Fig.5, the two straight lines defining "supply" correspond to eqn 8a with α_{max} set at arbitrary values of 1.2 or 2.4 (to represent a "restricted"

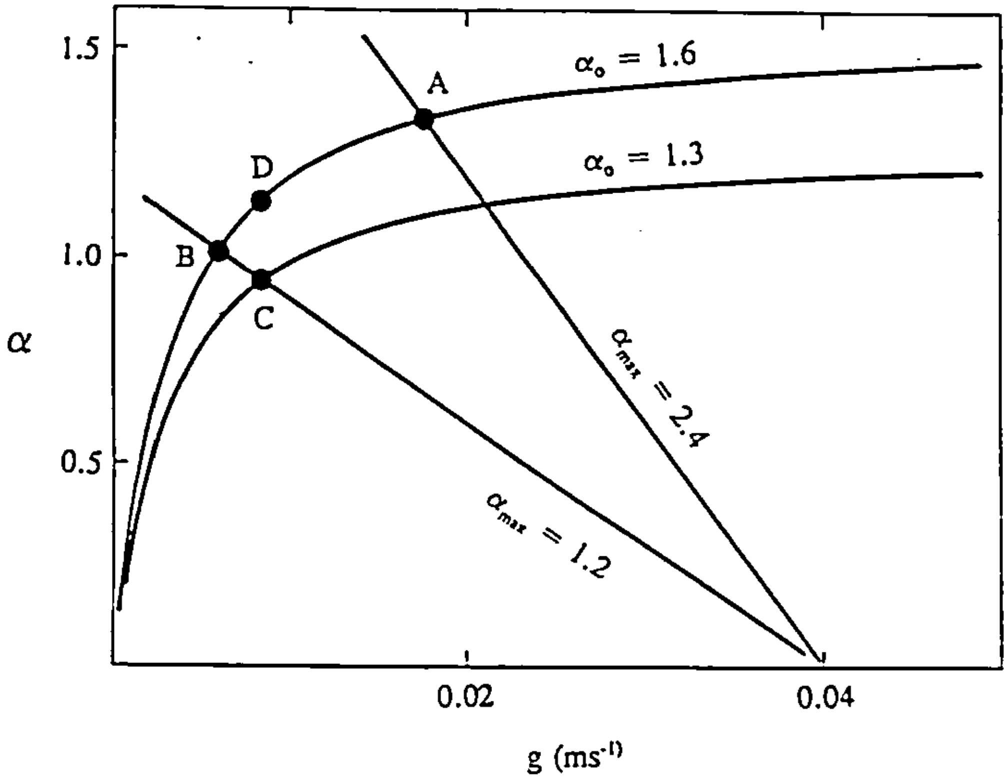

Fig.5 Dependence of Priestley Taylor coefficient α on (a) "demand" as specified by α_0 = 1.3 and 1.6 (hyperbolas); and (b) on "supply" as specified α_{max} = 1.2 and 2.4 (straight lines). For further details see text.

or an "unrestricted" supply of water for transpiration); and with g_{max} set at 0.04 ms^{-1} for both cases. (This is about twice the maximum conductance observed in the field (corresponding to r_s = 50 sm^{-1}) implying that maximum observed values of E will be about half E_{max}). The two hyperbolas represent equation 7 with extreme values of α_o (1.3 and 1.6) obtained from McNaughton and Sprigg's (1989) analysis of the Cabauw data.

Point A defines values of α and g when there is a maximum demand from the atmosphere (α_0 = 1.6) and a maximum supply (α_{max} = 2.4) by roots. In this case, PE(v) corresponds to α = 1.35 and the effective conductance of the canopy is 0.0174 ms^{-1} (r_s = 57 sm^{-1}). These values are characteristic of potential evaporation from well-watered vegetation completely covering the ground.

If, in drier soil, the supply is halved (α_{max} = 1.2), intersection of the new supply line with the upper hyperbola (B) implies a decrease of stomatal conductance by a factor of 3 (from 0.0174 to 0.0057 ms^{-1}). However, there is a much smaller decrease of transpiration rate corresponding to a decrease of α from 1.35 to 1.02, i.e. by about 24%.

The conservatism of α despite a major change in conductance is a consequence of physical feedback operating in two ways. Microclimatically, closure of stomata (by restricting water supply) is accompanied by an increase in leaf temperature and a decrease of vapour pressure in the ambient air. Both processes increase the gradient of vapour pressure between intercellular spaces and ambient air, partly offsetting the decrease of stomatal conductance. Macroclimatically, the CBL receives more sensible heat from the surface with the consequence that more warm, dry air is entrained at the inversion and this process also increases gradients of vapour pressure that drive the transpiration rate.

Conversely, suppose that when the supply of water is restricted ($\alpha_m = 1.2$), there is an increase in atmospheric demand equivalent to an increase in α_o from 1.3 to 1.6 (+23%). If conductance remained constant, α would increase from 0.90 (at C) to 1.13 (at D), i.e., by 26%. But because conductance is predicted to **decrease** from 0.0077 to 0.0057 ms^{-1} in response to the increased demand for water, the expected change is to $\alpha = 1.02$ (at B), an increase of only 13%. In this case, the decrease of conductance is a consequence of **physiological** feedback, probably mediated by gradients of water potential that determine transport between guard cells and mesophyll cells (Dewar,1994). Note that when the supply of water is "unrestricted" ($g \geq 0.04$ ms^{-1}), the impact of physiological feedback is negligible.

The conservatism of evaporation rates in this scheme partly explains why PE has often been successfully estimated from climatological measurements (e.g. using the Penman or Priestley-Taylor equations) without reference to biological constraints. The impact of relatively small changes either in "supply" or in "demand" will often be within the limits of observational error. Much more detailed descriptions of feedback including the role of CO_2 assimilation were given by McNaughton and Jarvis (1991) and by Jacobs (1994).

The next fifty years

The search for ways of estimating potential rates of evaporation and using them to derive rates of actual evaporation has made slow but steady progress over the past 50 years. Rates of potential evaporation can now be confidently estimated for extensive areas under more or less uniform arable crops or forest. Uncertainties remain about sub-potential rates of evaporation from crops (and implications for yield); and about evaporation from inhomogeneous ecosystems such as moorland or mixed woodland.

Side by side with the evolution of formulae described in this brief review, experimental methods of measuring evaporation have improved immensely. Cumbersome lysimeters measuring evaporation from somewhat unrepresentative patches of surface have been replaced by infra-red hygrometers and sonic anemometers with which accurate measurements of the mean flux of water vapour can be obtained, second by second, for days on end. This type of equipment has been deployed in major international campaigns in which the structure of the

Convective Boundary layer has also been explored. On a global scale, satellites provide continuous measurements of potentially usefully variables such as radiative surface temperature and microwave reflectivity.

In terms of data capture, new hardware has performed extremely well. The challenge now is to ensure that this wealth of information, undreamed of 50 years ago, will be fully exploited to test and improve the theoretical basis of evaporation science. I hope that the contents of this review will be completely overtaken by new developments long before potential evaporation celebrates its centenary in 2044!

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ADEQUACY OF THE IRISH RAINFALL NETWORK

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Abstract

The network for precipitation measurement is described and the adequacy of the information it provided during some severe events is discussed. Current Irish practice for mapping the precipitation field is outlined and a possible role for precipitation forecasts suggested. Finally, for redoing extreme rainfalls of various durations and return periods the merits of the log-logistic distribution are outlined.

The rainfall network

The Meteorological Service collects daily rainfall information from about 650 stations; this information is received monthly and is the single most important source of rainfall data. Of these 650 stations 169 were found to have an acceptable reading for each day of 1988 - a stringent requirement. A further 200 stations would be classed as highly reliable and most of the remainder are usable to some degree. Values back to 1941 are available on computer media. While there is considerable information on daily rainfall prior to 1941, it has not, as yet, been computerised. In lowland areas the coverage is adequate nearly everywhere but in mountainous areas, hydrologically the more interesting, the same cannot be said.

For falls of duration less than a day we rely on a network of Dines rain recorders (Fig. 1). Besides our thirteen synoptic stations, which report hourly falls, there are about 75 others. When tested in February 1990 only 19 were fully satisfactory, 22 needed minor adjustment and 35 were poor to useless. Unlike the synoptic stations, hourly values are not extracted routinely but the charts are stored. Thus, for hourly values we have the synoptic stations supplemented by the number of recorder stations which are working satisfactorily. At the synoptic stations we also tabulate the maximum falls for durations between 15 minutes and 24 hours, provided the falls exceed set thresholds e.g. 6 millimetres for a 60 minute fall. This data set has been accumulated since the late 50's.

Current Rain Recorder Stations
March 1992

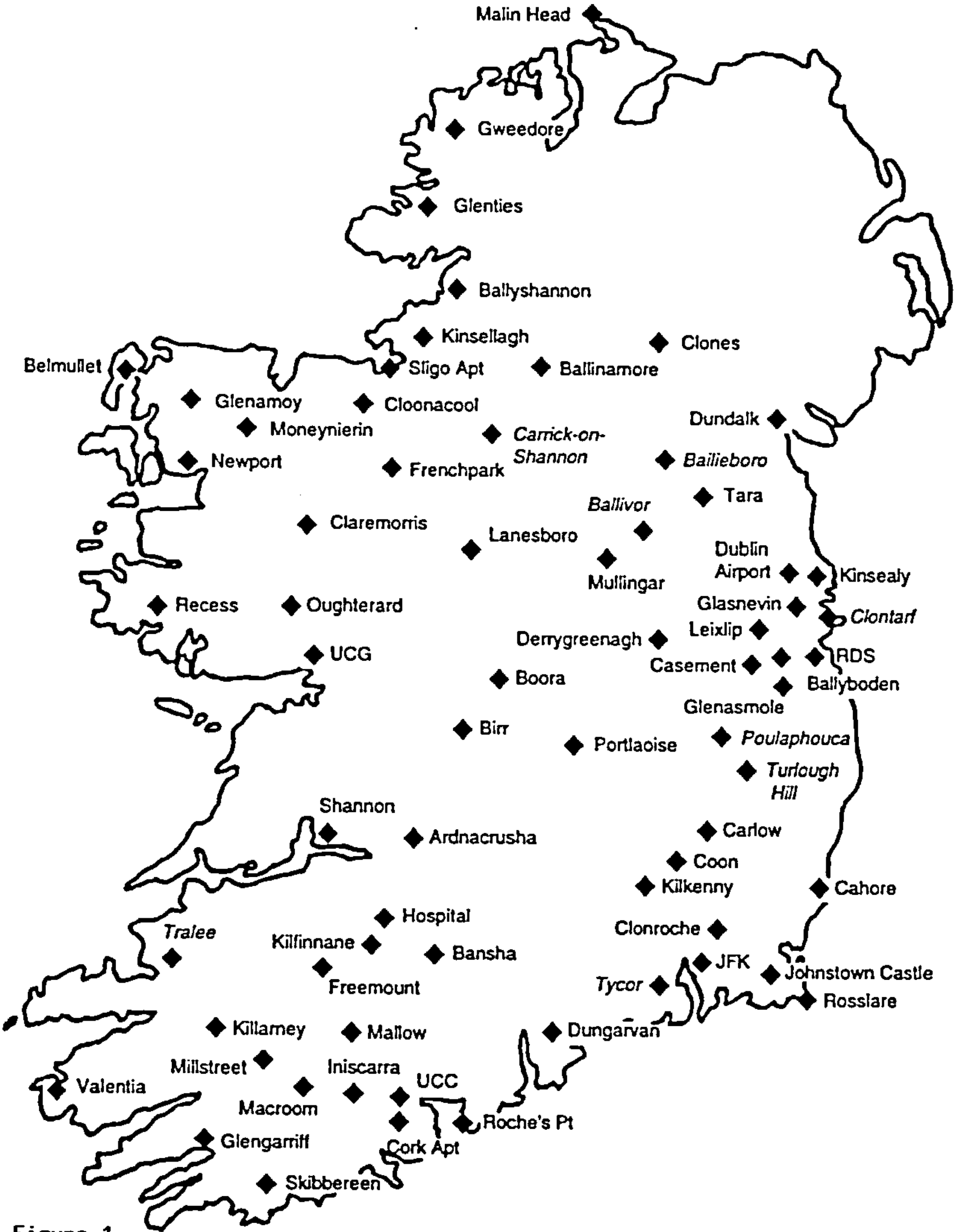


Figure 1

stations in italics have weekly clocks

Radar measurement of precipitation

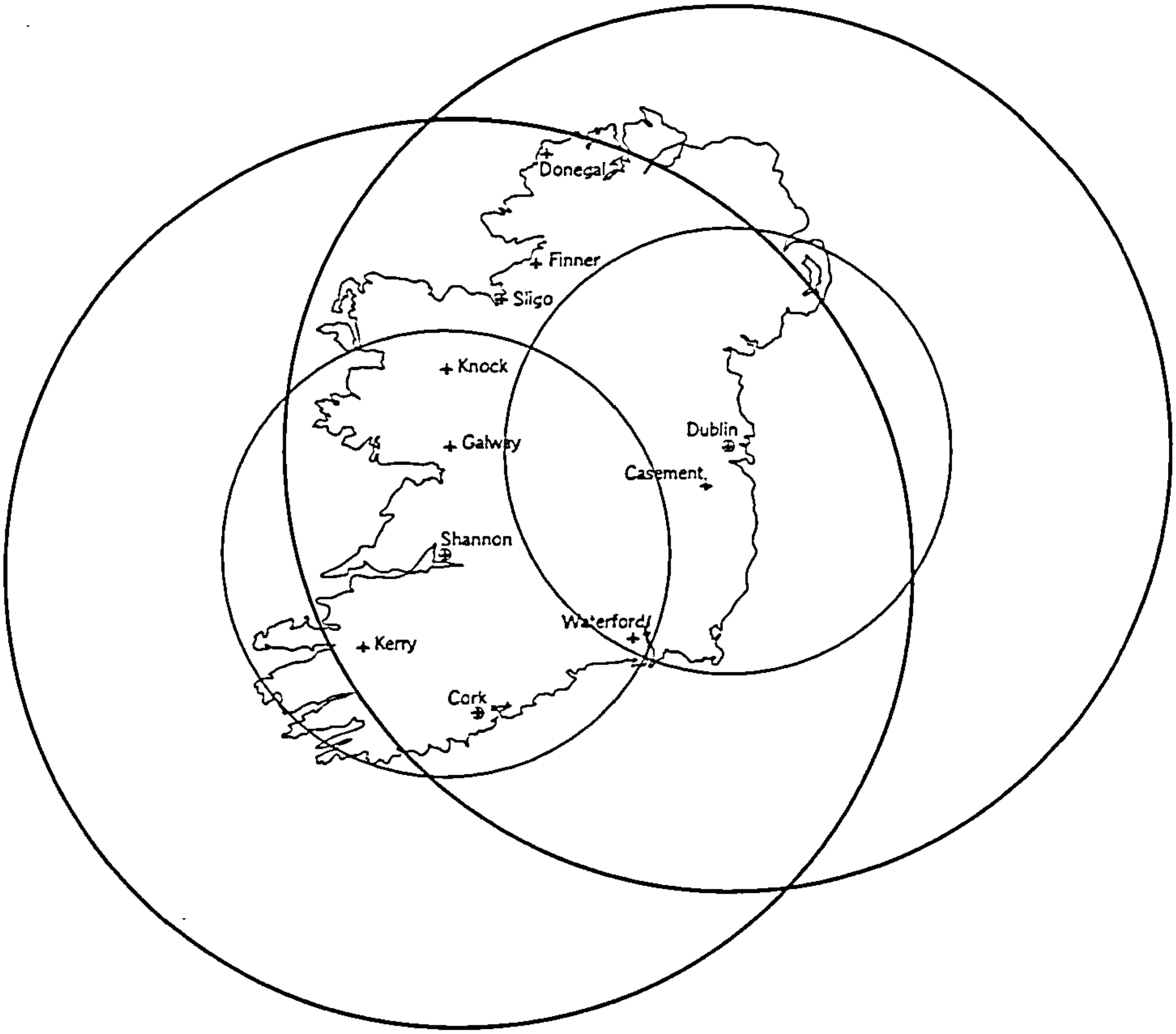
Coverage of the country is quite good (Fig. 2) as Ireland participates in the COST programme. To date radar has been employed mainly as an aid to forecasting but has been little used in climatology. However, our Ingres database has an extension in which images are stored as matrices and information can be extracted pixel by pixel. We have just undertaken a pilot study on the use of this facility but have, as yet, no results to report. We do feel that radar is important for the future, providing the hope of fairly accurate areal values.

Limitations of the precipitation network

The daily gauges are British Meteorological Office standard gauges and like any such device standing some 30 centimetres above ground level it is subject to losses caused by wind field deformation, rebounding raindrops and evaporation. Like many other agencies involved in rainfall measurement we have conducted experiments comparing the catches of pit gauges, ordinary gauges surrounded by turf walls and gauges exposed in the normal manner. The results were, unsurprisingly, that pit gauges received monthly totals usually between 5% and 15% higher than a gauge standing in the open while the gauges with turf walls recorded totals considerably closer to the pit gauge. For gauges with ostensibly normal exposures there can be significant differences over a small area. At our Malin Head station the difference in monthly totals between gauges in front and behind the station averaged nearly 8% over a period of years, mainly because of wind effects. Watkins (1955) put 9 gauges in an area of 60 square feet and over a period of 21 months found that the maximum difference between the gauges was 4 to 5% but in most months was about 2%. During 5 heavy falls of rain the maximum difference was nearly 8%. All this accumulated uncertainty about the accuracy of point rainfall and its conversion to surface rainfall adds to the problems of the hydrologist.

Rain recorders differ from the standard gauge in height above ground and in aperture size; also there is the loss of rain when siphoning takes place. These differences are accounted for by correcting the recorder total to agree with the gauge and the correction is usually fairly small. The crowding of the traces on the chart during

Map of Ireland showing coverage of the Dublin and Shannon radars



inner circle: range of accurate data and all wind information
outer circle: Extended range of the radar with good coverage of the location and movement of precipitation

Figure 2

periods of heavy rain can be a problem but again the difficulty can usually be resolved. However, experienced observers used to tabulating Dines recorder charts think 15 minutes a reasonable lower limit of resolution and feel distinctly uneasy with any period less than 5 or 6 minutes. We have the more sensitive, faster-response Jardi recorder at a few stations and the rates of rainfall they give for 5 minutes are invariably higher than the Dines, indicating that the instrument used for recording short-period rainfall should be known to anyone using the records.

The radar at Dublin Airport has a range of about 240 kilometres but is effective for rainfall measurement only to about half that distance. In a test last year the coefficient of determination for Mullingar, 75km to the west, was 0.75 while for Clones, 107km to the north, it reduced to 0.62. The scatter was large (Fig. 3) indicating that while a generally useful indicator of point rainfall, it needs to be used with great caution. A recent British study (Cheng and Brown, 1993) indicates that areal rainfall rates can be found to an accuracy of 14% to 20%, while 5-day totals can be found with an average error of 8% to 11% when the rain is mainly frontal in origin. Such accuracy achieved operationally would be very good news indeed for the hydrological community.

Performance of the Network

On 11th June 1963 we had a small-scale severe thunderstorm in Dublin City. With more than 20 gauges in the general area it might reasonably be expected that coverage would be adequate. The highest total recorded was 98mm at Ballsbridge with about 10mm at Dunlaoghaire, 40mm at Clontarf and 2mm at Howth (Fig. 4). An appeal was made through the newspapers for any readings which had been made by private individuals and among the replies received were several from the Mount Merrion, Sandymount and Churchtown areas which indicated totals of 150 to 185mm; one of these proved to be from a well-exposed standard gauge whose owner had emptied the bottle several times during the storm and had recorded 83mm in 65 minutes. The highest hourly fall recorded by the official network was 46mm at Ballsbridge. It is sobering to realise that in one of the most densely-gauged areas in the country our regular network would not have provided good information on this small-scale thunderstorm. Curiously enough, on 11th June 1993 there were again very heavy falls

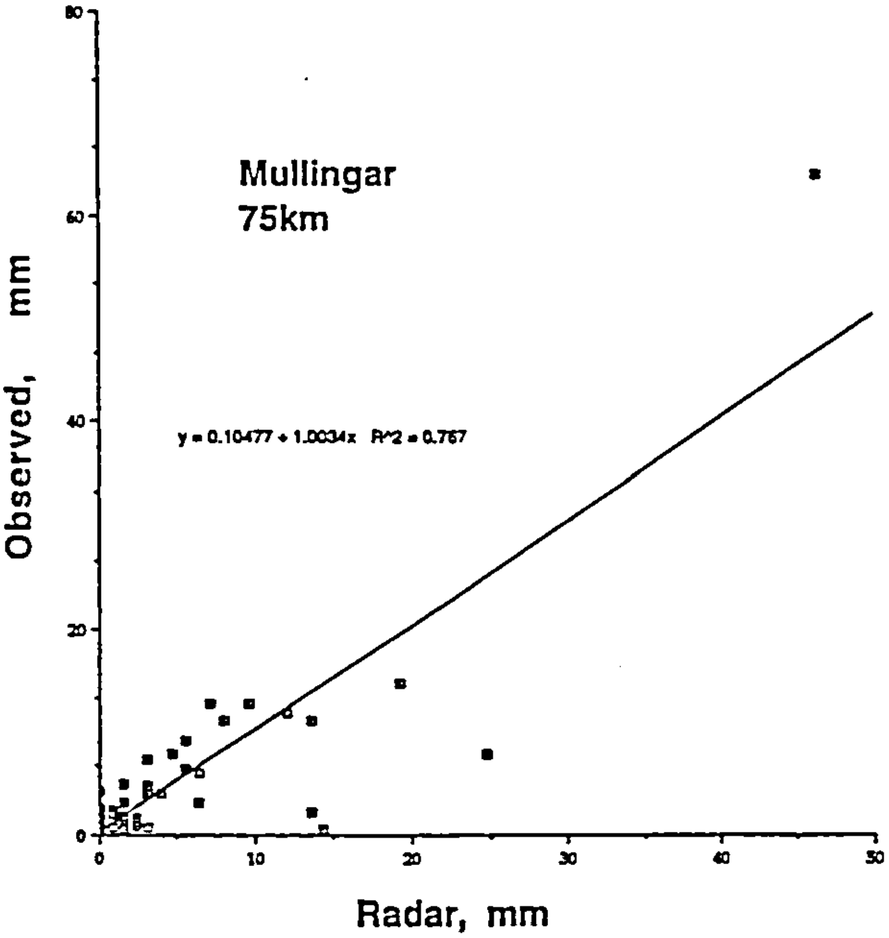
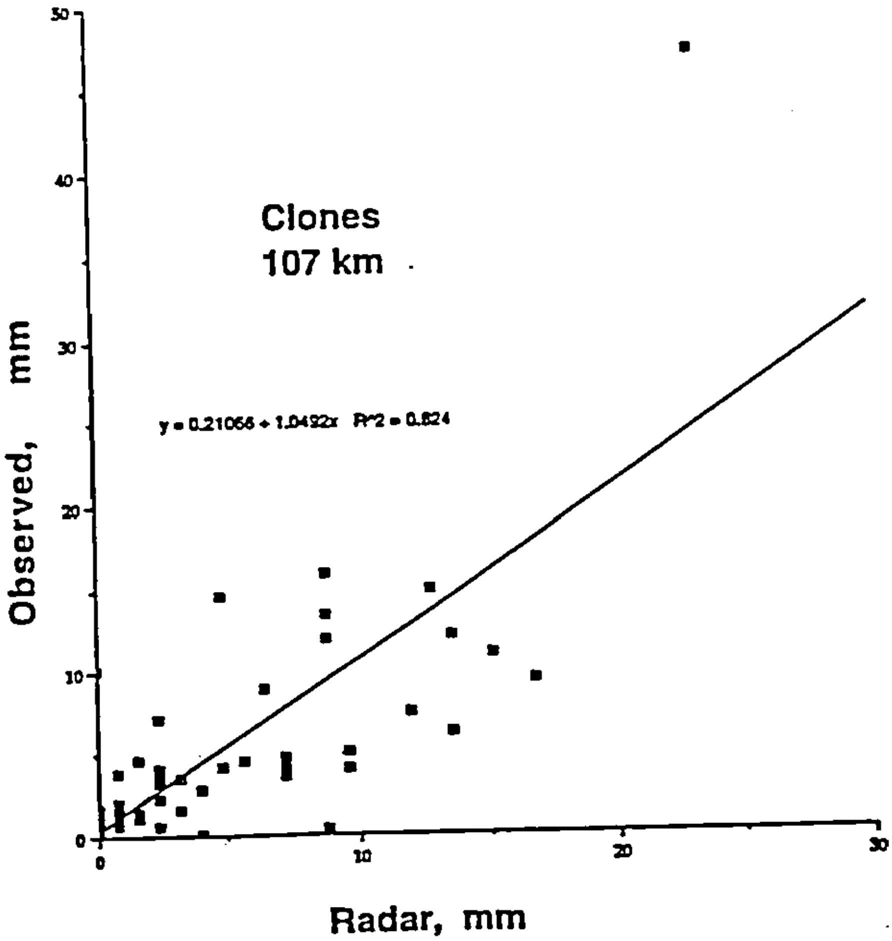


Figure 3. Observed daily rainfall amounts compared to radar estimates at Mullingar and Clones (below).



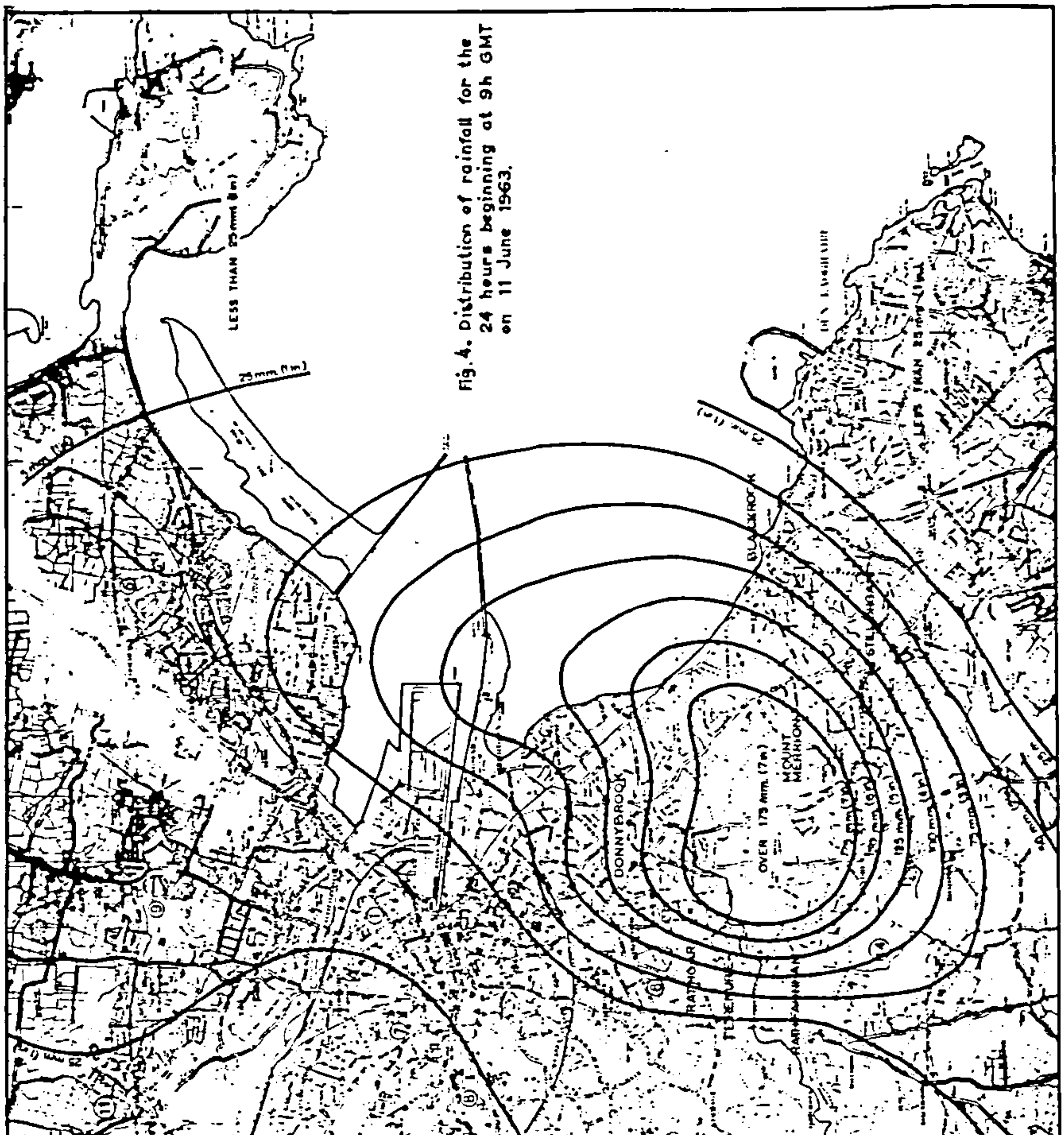


Fig. 4. Distribution of rainfall for the 24 hours beginning at 9h GMT on 11 June 1963.

of rain in the Dublin area but they were larger-scale and less variable and the network provided a good picture of the distribution of totals which were over 100mm in places.

Down in the southwest 18/19 September 1993 provided a surprise. In what appeared an unexceptional situation - an active front approaching slowly from the Atlantic - the falls were very variable and included the highest fall on record in the southwest with Cloone Lake recording 243.5mm; this exceeded by 37mm the falls recorded in the area during the very severe storm of 5/6 August 1986 when a deepening and extremely active wave depression forming in the south Atlantic moved right across the southwest causing considerable flood damage. The situation of September last appeared much less active but the network did provide supporting evidence with falls of 95 to 175mm recorded at other stations. The Shannon Airport radar, while restricted by the mountainous terrain, did indicate small areas of intense activity on the front. The performance of the network for this event was satisfactory.

Areal estimates of rainfall

Many and varied have been the schemes devised to estimate rainfall over an area from point values measured in the vicinity. In the WMO report of Liebscher (1993) it is asserted that 'none of the mathematical methods is able to replace interpretation by an experienced climatologist'. In the Irish Meteorological Service, a surface fitting method (Hamilton et al., 1987) is used; it has been devised so that it gives results very similar to the manually drawn maps. Percentage maps proved relatively easy to mimic. Monthly rainfall amounts were much more troublesome as the results in mountainous areas were not satisfactory. The solution found was (1) to use closed stations for which long-term averages were available and to calculate amounts using the percentage field (2) insert dummy values in a few crucial areas which never had a station - the average amounts were read from the (manual) maps and the percentage field used. The method requires no manual intervention and, in part, contains an attempt to translate the experience gained over many years of drawing rainfall maps into what might be termed an expert system. We find the method quite satisfactory

(Fig. 5) but in a way that amounts to no more than saying that it enshrines our prejudices. In the future we should get more objective areal values using estimates based on radar.

A point of interest here is that weather-forecast models now produce estimates of rainfall. Forecasters, who have to advise on the amount of rainfall to be expected, find these fields a useful rough guide but experience of their limitations is necessary. As computing power increases enabling the mesh size to be decreased and more realistic parameterisations of the thermodynamical processes involved, we may anticipate considerable improvement in these forecast precipitation fields.

Design Values of Rainfall

Rainfall amounts for various durations and return periods are among the most frequently requested types of hydrological information. The tables we use at present are based on the FSR (Flood, Studies Report, 1975) and there is a general feeling in the hydrological community that it needs to be redone. The FSR fitted annual maxima to a generalised extreme value distribution. Since 1975 we have acquired much more data and there has been much work on the analysis of extremes but no general agreement has emerged. Cunnane (1989), in a comprehensive review, lists 15 distributions which have been extensively used; he also stresses that the method of estimation should be specified as well as the probability distribution as this has important consequences for quantile estimation. He eventually plumps for the Wakeby distribution with the probability weighted moment method of estimation. The Wakeby distribution is a 5-parameter brute of form:

$$x = m + a\ (1-(1-F(x))^{b}) - c\,(1-(1-F(x))^{-d})$$

and is not meant for single sites but rather for regional analysis; this requires selecting a set of stations which are regarded as possessing the same distribution when the data at each station had been suitably scaled, usually by dividing by the mean. For rainfall, selecting a region is less of a problem than for floods. Cunnane (1989) refers to the 3-parameter log-logistic distribution as put forward by Ahmed et al. (1988). Canadian workers (Shoukri et al, 1988) found the 2-parameter version suitable for maximum

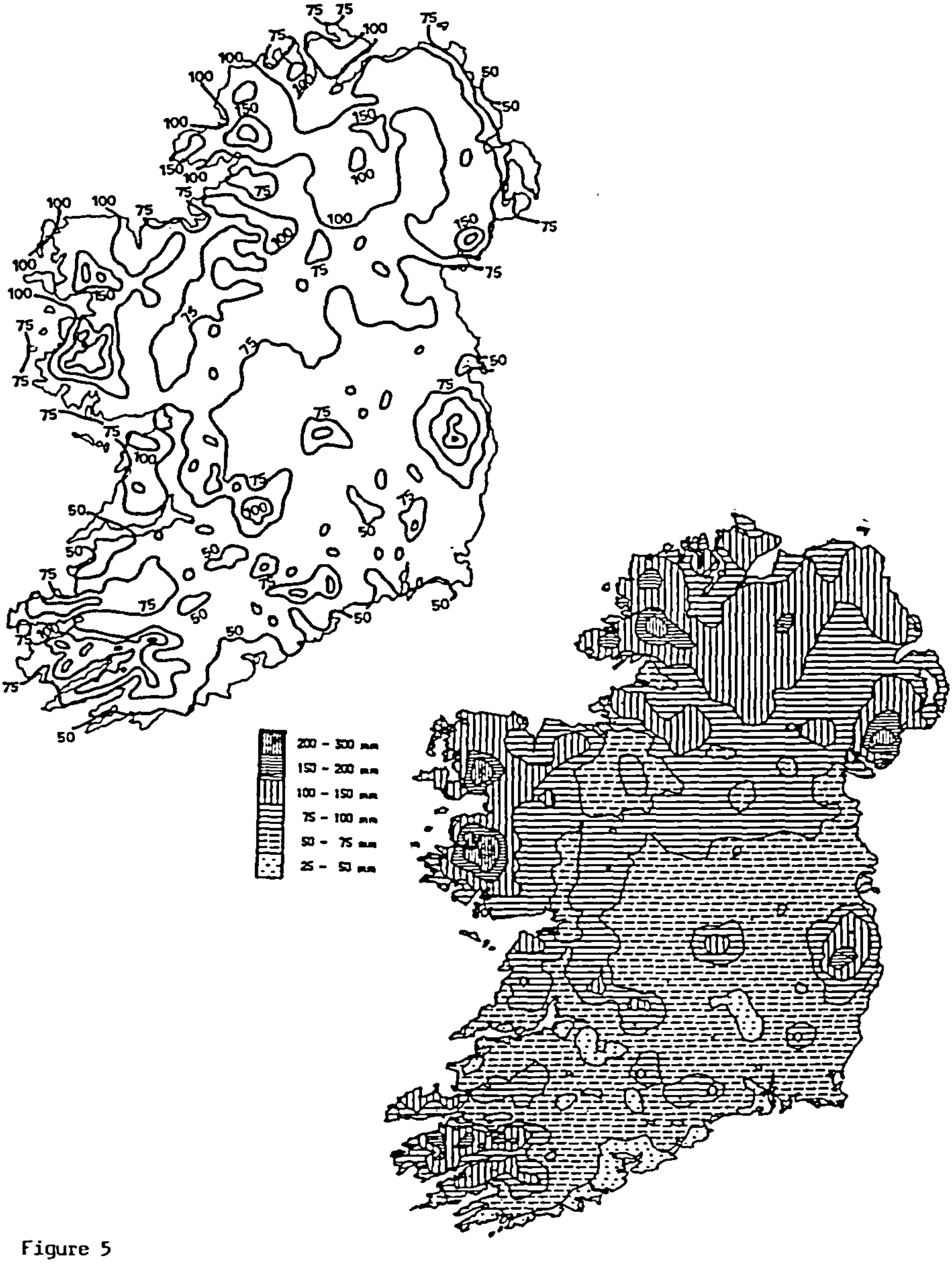


Figure 5

Comparison of manually drawn isohyet map for April 1986 (top) with automatic plot

daily falls; as an exceedance distribution this has been the subject of some preliminary study in the Irish Meteorological Service.

An alternative to using annual maxima is to use the peak over threshold (POT) method, with a Poisson process of occurrences described by a rate parameter (r), and a two-parameter distribution for the exceedances which are taken as independent of the occurrence process; Madsen et al (1993) describe this type of model with the generalised Pareto distribution (GPD) for the exceedances. The GPD can be derived as the limit of high exceedances over a high threshold and is a degenerate Wakeby distribution with $y = a(1-(1-F)^b)$. Typical of the behaviour of distributions with a shape parameter is that quantile estimates are very sensitive to the value of the shape parameter and its variance; this is true for the GPD for which the value estimated is also very dependent on the top values in the sample. The 2 p Llg distribution is suitable for use as an exceedance distribution and exhibits some nice characteristics. The form of the distribution is $x = a\,(F/(1-F))^c$ where a is the median, F the cumulative distribution function and c the shape parameter. For a process averaging r exceedances per year, the growth curve of x_T, the value exceeded on average once in T years is $x_T = a\,(rT-1)^c$ and the shape parameter is the exponent of this power law curve which is very similar in form to the curves determined empirically for the FSR. The model is in terms of 3 parameters (r,a, c) which have a clear interpretation and this sort of formulation can be useful when considering climate change scenarios. The maximum likelihood estimators of the parameters and quantiles are good for moderate sample sizes. The crucial shape parameter has a dependence on very high and very low values according to the logarithm of the values - it is thus relatively insensitive to changes in the top values which is a valuable characteristic but care must be taken that low values are not too influential. In summary then, the log-logistic distributions seem very promising whether used for annual maxima or exceedances. However, it is just as well to recall the observation of Moran (1957) on the subject of fitting curves and extrapolating the tails: 'the form of the distribution is not known ... the part of the distribution we are interested in is well away from the part where observations provide some information ... this cannot be overcome by any mathematical slight of hand'.

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RIVER FLOW IN IRELAND - CARACTERISTICS AND MEASUREMENTS

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Abstract

A short account of the development of river flow gauging in Ireland is presented and methods currently used for such gauging are described briefly. Average Irish river flow characteristics as reflected by mean runoff, flood magnitudes, low flows and drought volumes and evaporative loss are outlined in numerical and graphical form.

Introduction

Ireland is saucer shaped with a high maritime rim and a flat low central plane. As a result of this configuration, many rivers that have their sources on the inland side of maritime mountains follow long roundabout courses to the sea with consequent sluggish flows. In times of heavy or prolonged rainfall such slow moving rivers are incapable of conveying the water which can result in serious flooding. It is this flooding in particular which has prompted the growth of Irish engineering hydrology during the last century and a half.

Here we examine numerically the general characteristics of river flow in Ireland.

Surface Water Measurement

Surface water measurement in the Republic of Ireland is carried out by three organisations:

(a) Office of Public Works (OPW)

(b) Electricity Supply Board (ESB)

(c) Local Authorities and Environmental Protection Agency (EPA)

These organisations began their stream gauging activities at different times and with different objectives.

The Office of Public Works has been involved in flow gauging for over 100 years in connection with its activities in the field of arterial drainage. In 1939, hydrometric survey was undertaken in a systematic and comprehensive way and since the early 1950's most gauging stations have been equipped with autographic water level recorders.

The original and primary purpose of the OPW. gauging stations was to gather information on flood flows for use in the design of arterial drainage schemes, so most of them were situated on medium to large rivers in areas needing drainage. Many are now used to monitor the effects of completed schemes. Despite this specialised purpose, there has been, since the beginning, an interest in low flows as well, although in come cases the physical properties of the stations were not wholly satisfactory for low flow gauging. In recent years reliability and precision of rating curves at low levels have been improved by more frequent gauging and, in some cases by resiting the station or by constructing artificial controls.

The Electricity Supply Board which was formed in 1927, began river gauging about that time on a small scale on the river Shannon. Subsequently, in the 1930's a network of gauges was set up on rivers which had some potential for hydroelectric development. This network was later expanded as interest in hydroelectric power sharpened due to the energy crisis. The ESB. has now decided on its requirements and has transferred some stations to OPW and local authorities and abandoned others.

While some Local Authorities had earlier maintained records of flows in rivers that were the sources of municipal water supplies, these gauging stations did not form part of a network in any strict sense. The Water Pollution Act (1977) placed considerable increased responsibilities on Local Authorities for the management and planning of the use of surface water in their areas with particular reference to abstractions of water from river systems and the emission of wastes. A major river gauging programme was undertaken for these purposes under the direction of An Foras Forbatha (AF), whose Water Resources Division had been actively engaged in hydrometric work since 1972. The duties of the water resources division were transferred to the Environmental Research Unit (ERU) in the late 1980's and in 1993 they were incorporated into the newly formed Environmental Protection Agency. The EPA has statutory responsibility for preparing a National Hydrometric Programme and discussions are proceeding with relevant authorities to prepare a comprehensive programme.

Gauging Station Network

There are over 1200 gauging locations in Ireland, approximately 500 of which have automatic recorders, - 400 on rivers and 100 on lakes and estuaries.

The total hydrometric archive includes more than 15,000 gauge years of water level data, over 40,000 current meter measurements and up to 5,000 gauge years of processed flow data. The data available include continuous water level measurements, mean daily flows, monthly and annual flow values and various statistics (minima, means, maxima, drought volumes) derived from them. In addition, the output from analysing such data, without involving any statistical assumptions, are also frequently available from the gauging authorities viz. flow duration curve data, sustained flow data, monthly maxima and minima, and drought deficits relative to stated demand levels. In addition, current meter measurement data are also available.

A register of all hydrometric gauging stations in the Republic of Ireland is maintained by EPA. All gauging stations in a particular catchment area share the same two leading digits in their station number. e.g. 0709 and 0714 are stations within the Boyne catchment.

River Flow Measurement

This consists of water level measurement at a gauging site and determination of the stage or water level - discharge relationship (rating curve) for the section. In earlier years water level was read visually by a local observer, once or twice daily, whereas at present water level is recorded on a chart by an automatic recorder. The recorder is operated by a float mechanism installed either in a stilling well or in a 300 mm diameter access tube. In a few cases a bubble mechanism is used instead.

The rating curve is generally estimated by current meter measurements. In the case of fifteen sites Crump weirs or flat Vee Crump profile weirs have been installed. In the latter case the theoretical rating is checked by reference to current meter measurements up to the modular limit of the weir and is rated by current meter measurements above that limit.

The measurement is carried out using current meters and flows are estimated by velocity area methods. The standards used are the international standards for flow measurement adapted to suit the particular gauging locations based on analysis of the long records of measurements at the site and the experience of measuring flows in Irish conditions e.g. at certain sites, on flashy rivers, modified measurement procedures are used to give the best estimate of the flow during the short time that the floods are near their highest level.

Other methods of river flow measurements, such as ultrasonic or magnetic methods which have been developed and demonstrated elsewhere have not thus far been employed in Ireland.

In ideal circumstances, such as near the upstream end of a long uniform stable channel reach, a unique rating curve could be developed from current meter measurements that would be valid for many years to come. The only limitations on its accuracy would be those due to the measurements themselves or in the instruments or techniques used to determine them. In many cases in practice this uniqueness and stability is unfortunately not evident despite the best efforts of those personnel involved. Such variability in time between measurements of flow, for the same observed water level, occur because of variations in "control" caused by weed growth and/or changes in the river bed sediments, principally. This is a principle that should be understood by all users of hydrometric data, especially data relating to low flows where it may appear difficult to reconcile individual current meter measurements with published flow data.

Some examples of such variations in rating curves are shown in Figure 1. These show the rating curves from a weir, a relatively stable site and a site where there is considerable seasonal variation. It should be noted that much of this variation can be accounted for by chronological and seasonal changes at the site and when suitable adjustments have been made the resulting scatter is within acceptable limits and separate rating curves are constructed for different periods of the year.

Characteristics of Irish River Flows

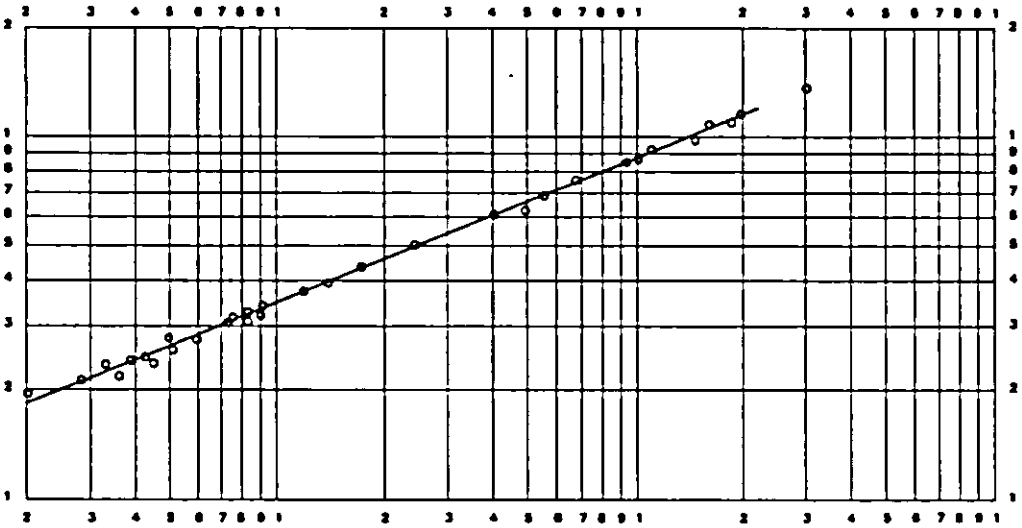
In general, Irish rainfall amounts are generous and evaporation amounts are not excessive so that the average specific runoff is of the order of $2.0 m^3/s/100 \ km^2$. This specific runoff varies from approximately $3 \ m^3/s/100 \ km^2$ along the western seaboard to approximately $1.5 \ m^3/s/100 \ km^2$ in the eastern part of the country, see Figure 2. Higher and lower values can of course also be observed.

In current practice flood peaks, Q_T, of return period T years, are often expressed as

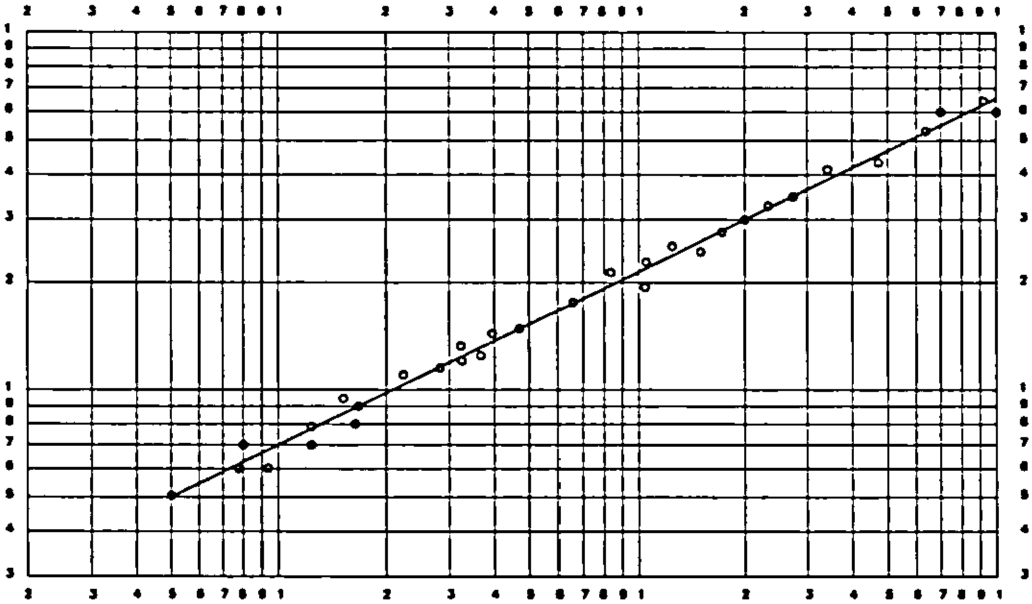
$$Q_T = \bar{Q} \cdot X_T$$

where $\bar{Q}$ = mean of the annual maximum flood series for any catchment and

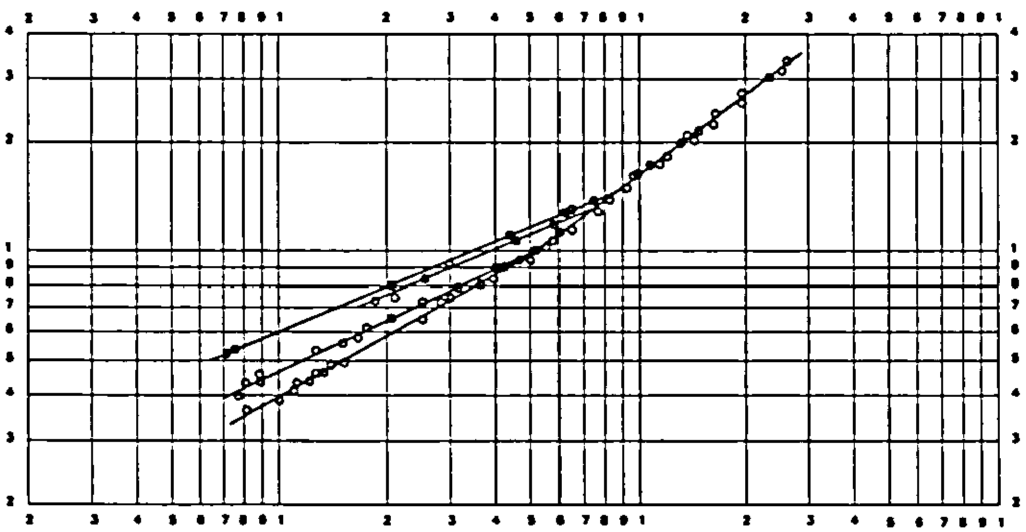
$X_T = Q_T / \bar{Q}$ is the growth factor relating Q_T and $\bar{Q}$



(a) Weir.



(b) Relatively stable site.



(c) Site affected by vegetation growth etc.

Fig. 1. Examples of Rating Curves.

IRELAND

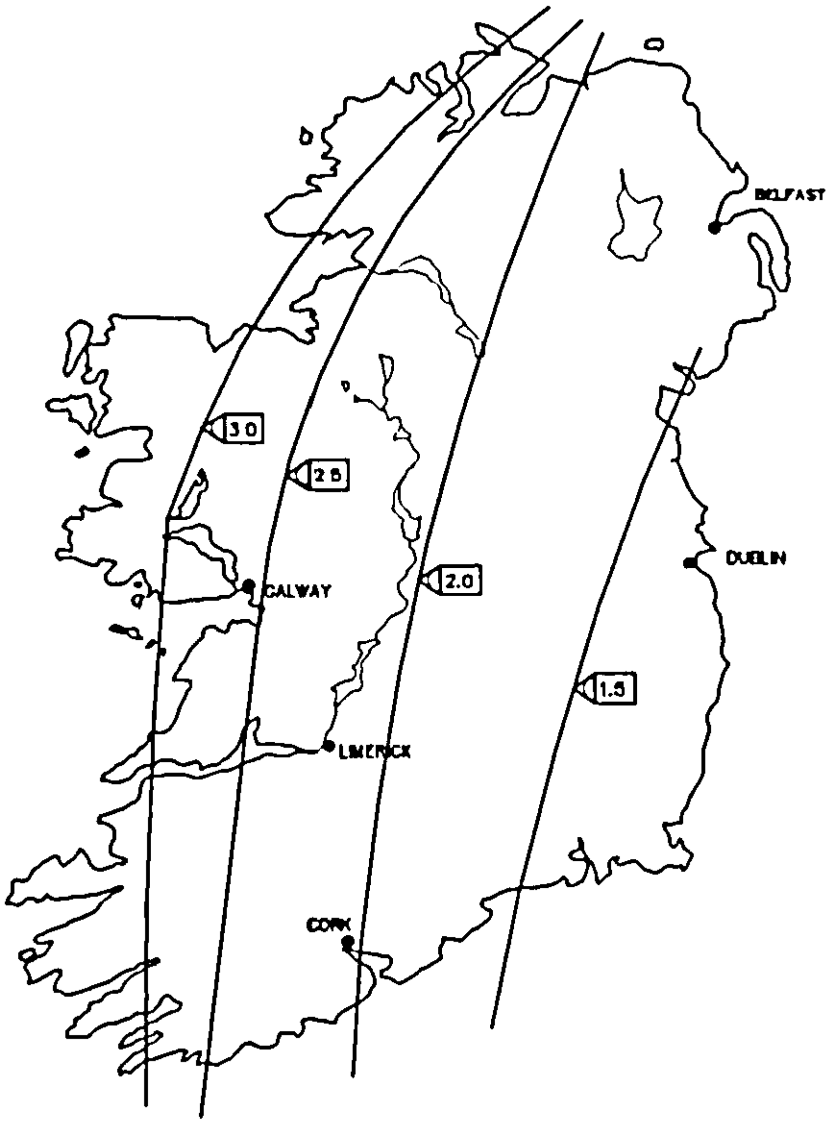


Fig. 2. Average Specific Runoff $m^3/sec/100km^2$

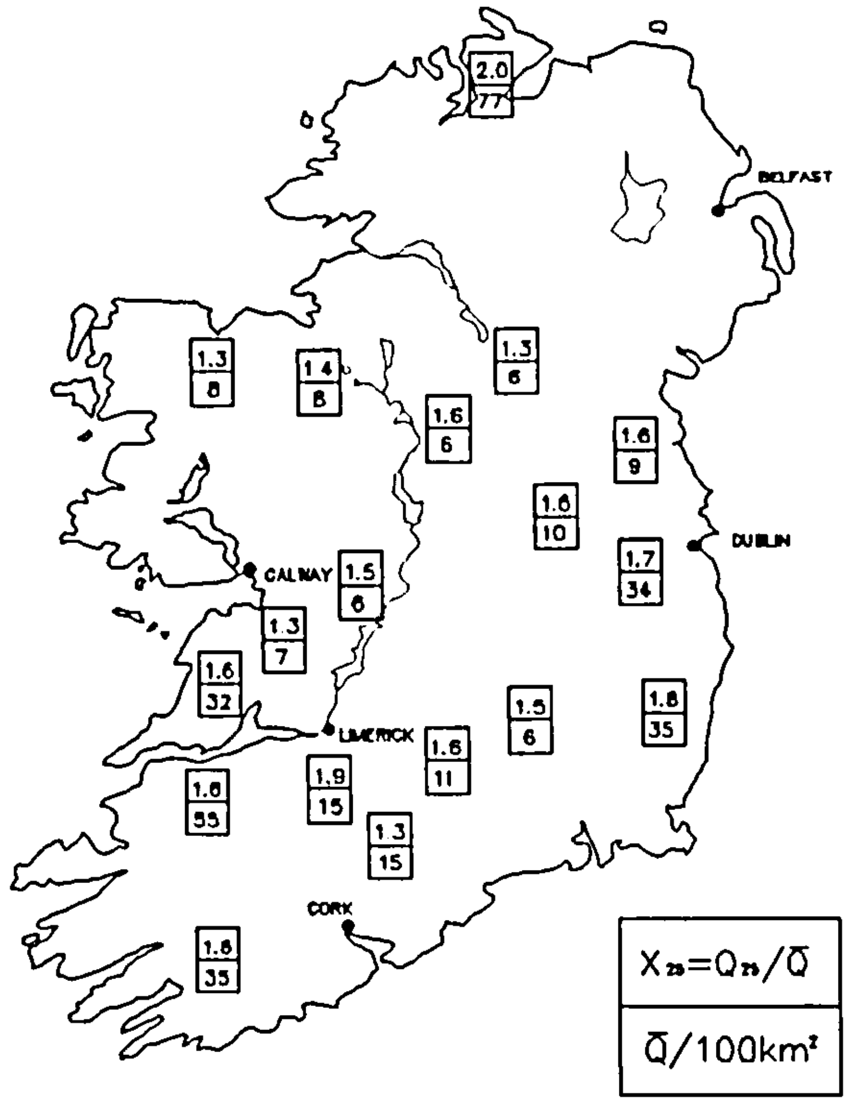


Fig. 3. Values of Flood Growth Curve for T=25years and values of mean annual Flood per $100km^2$

Flood estimation for ungauged catchments and catchments with short periods of records use a regional average value of X_T with $\overline{Q}$ obtained from a catchment characteristics formulae or from the short record as appropriate.

For example, for T = 25 the value of X_{25} obtained for Ireland by the Flood Studies Report (NERC, 1975) is 1.60. Values of X_{25}, estimated from observed records at 32 gauging stations have an average of 1.56. The values from 18 of these are shown in Figure 3 while values of specific mean annual flood in m^3/s per 100 km^2 are also shown. These show the difficulty in predicting from individual catchment characteristics, in the sense that the variable area (A) does not explain the variance in $\overline{Q}$. Histograms of these quantities from the 32 stations are shown in Figure 4.

Data on low flows and especially extreme low flows are less easy to generalise as specific low flows (q/catchment area) tend to vary very widely depending not only on climate, vegetation and topography but also on soils and geology. In the series of annual minimum flows q_{50} is the value below which flow drops in one year out of 50. Martin and Cunnane (1977) found that a lower limit for q_{50} as a percentage of mean annual runoff $\overline{F}$ can be expressed as

$$\text{Min. } q_{50} = 2\% \, \overline{F} \qquad \text{when } A < 800 \text{ sq. km.}$$
$$= [2\% + 0.5\,(A - 800)/100\ \%] \, . \, \overline{F} \qquad \text{when } A > 800 \text{ sq.km.}$$

One method of quantifying droughts is to calculate the volume of water needed to be in storage prior to the commencement of the drought in order to augment the flow to the required demand level. Depending on the required usage (water supply, dilution, fishery requirements ...) the demand flow rate could range from 10% to 30% of the long term mean low. As an example, the record at station 3619, River Erne at Belturbet was examined and drought volumes calculated for demand levels of 50%, 40% 30%, 20% and 10% of the long term mean flow. These are shown in Table 1.

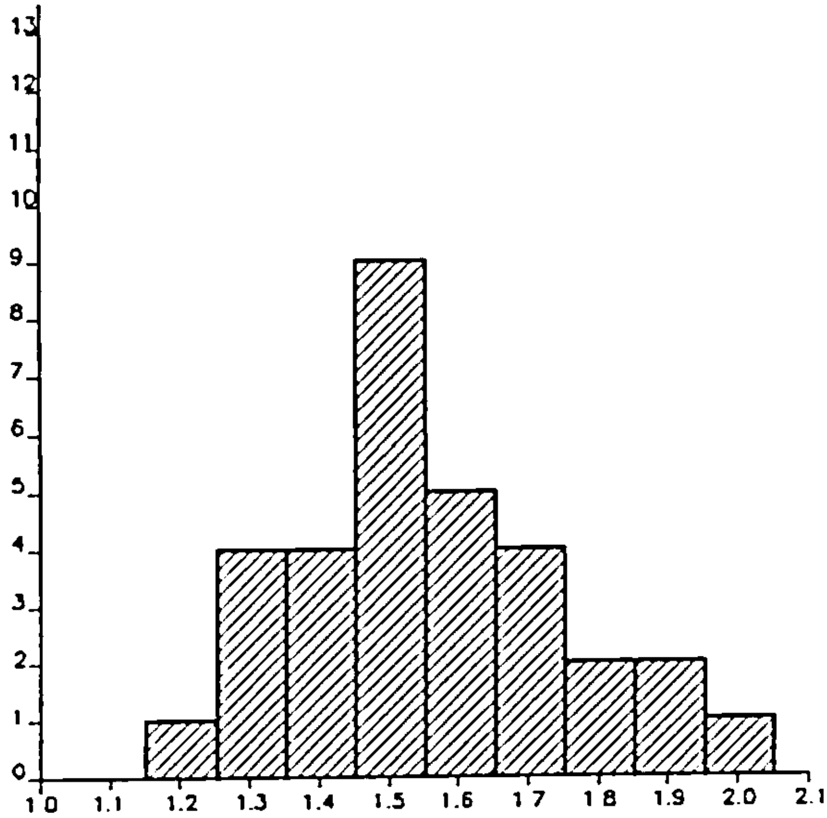


Fig. 4(a) $X_{25}=Q_{25}/\overline{Q}$ for 32 Stations

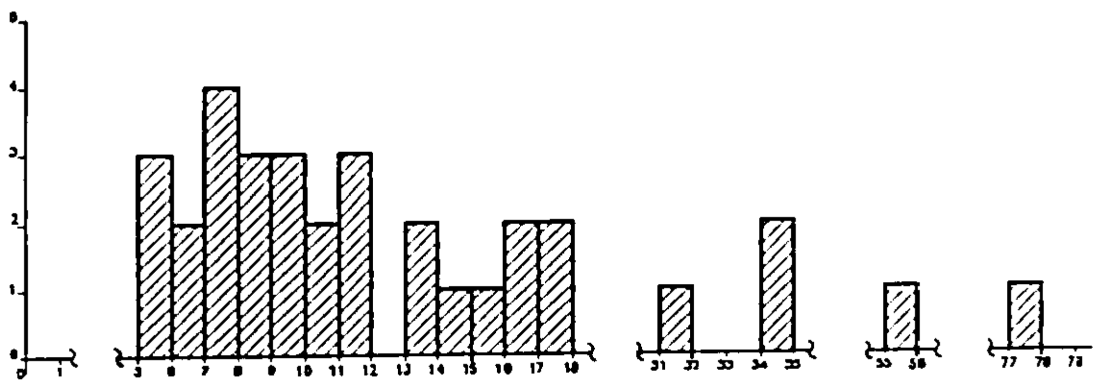


Fig. 4(b) $\overline{Q}/100km^2$ for 32 Stations

Table 1 River Erne at Belturbet - Estimates of drought volumes (in cumec days) of various return periods (1 cumec day = 86400m^3).

DEMAND LEVEL	RETURN PERIOD			
%MAF	2	5	10	25
50	1010	1470	1780	2460
40	690	1050	1290	1590
30	410	660	830	1040
20	180	340	440	570
10	27	90	132	185

Such analyses have been conducted by a number of researchers during the past two decades (Martin and Cunnane, 1977, Smyth 1984, Kachroo 1992, Martin 1992, Mac Carthaigh 1992).

Attempts have also been made to generalise these results so that relations between dimensionless volumes (S/V)% and their exceedance probabilities could be simplified so that (S/V)% might be estimated for planning purposes.

Water Balance

The long term evaporative loss can be obtained from the difference between mean annual rainfall and mean annual runoff expressed as a depth. Published studies (e.g. Martin & Cunnane 1977) and later unpublished studies indicate that this difference generally lies in the range 400 mm/year to 600 mm/year with an average value of 475 mm/year. Occasional values lie outside this range.

Conclusions

Ireland's saucer shaped topography leads to sluggish rivers in the interior plains and consequent flooding problems. An extensive river and lake gauging network has been developed over the past 60 years. The hydrometric archive now includes more than 15,000 gauge years of water level, over 40,000 current meter measurements and up to 5,000 gauge years of processed flow data. Average specific runoff is of the order of 2 $m^3/sec/100km^2$ and the annual evaporative loss averages approximately 475 mm/year. Flood magnitude return period relationships do not display a steep growth curve - the value of $Q_{25}/\bar{Q}$ is 1.60 which is typical oft the humid zone. Low flows are less easy to generalise except to say that q_{50} = 2% (mean flow) in small catchments.

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SIMILARITY AND SCALE EFFECTS IN THE WATER BALANCE OF HETEROGENEOUS AREAS

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Abstract

There has been a great deal of recent research activity concerned with macroscale hydrological modelling for the prediction of both runoff and land surface to atmosphere fluxes. Different approaches to this problem are discussed in the context of hydrological similarity within the complex heterogeneity that is the real hydrological system at any scale. A disaggregation approach to macroscale parameterisation is advocated, that is allowed to reflect the distribution of smaller scale responses. Some example calculations are given to demonstrate the concept of functional hydrological similarity of parameter sets and ideas for an approach to use this effective similarity in macroscale calculations are outlined. It is suggested that the identification problem can be reduced to the determination of a set of weighting coefficients for the index parameter sets.

On Hydrological Similarity in the Water Balance of Heterogeneous Areas

The water balance equation is a simple expression of mass continuity: inputs equal outputs plus change in storage. As such, it is the one equation in hydrology that scales directly; in principle, it can be applied at all scales from the point to the GCM grid square. Application of the water balance equation at either "point", catchment or macroscale, however, belies this simplicity, primarily due to the heterogeneity in the components of the water balance and the lack of direct measurement techniques for some important components.

However, in many applications it may not be necessary to have a complete assessment of the detailed water balance everywhere in the landscape. What is commonly required is a simple partitioning of the output components of the water balance into evapotranspiration, rapid (surface and subsurface) and delayed (subsurface) discharges. This partitioning is, of course, dependent on the heterogeneity and dynamic responses associated with individual locations and yet the range of hydrological responses within any physioclimatic region will be limited, being conditioned by the precipitation inputs and meteorological forcing for that region. Thus it may be that many points within an area may be hydrologically similar in their responses and water balance partitioning even though they are very different in their soil, vegetation and hillslope characteristics. If this similarity can be assessed then macroscale hydrological predictions might be greatly simplified.

There has been very little work on the problem of hydrological similarity in the past, particularly in the partitioning of the water balance. The work on the Geomorphological Unit Hydrograph (reviewed by Rodriguez-Iturbe, 1993) that was induced by trying to account for drainage basin topology and morphology was concerned primarily with runoff routing and not with runoff production and so will not be considered further here. Early approaches concerned with runoff production were catchment scale approaches in which catchment characteristics in the form of indices were incorporated into multiple regression analyses to produce predictive equations for runoff given a certain rainfall (a classic example is the UK Flood Studies Report, NERC, 1975). There have been no similar studies of evapotranspiration fluxes, where nearly all the available predictive equations are primarily driven by meteorological variables. Soil-Vegetation-Atmosphere-Transfer (SVAT) models do take into account soil and vegetation characteristics but until recently their application at larger scales has been simply a matter of taking a local SVAT model and multiplying by the area of interest (such as in the application of the SiB (Sellars et al., 1986) and BATS (Dickinson and Kennedy, 1991) SVAT models at the GCM grid scale).

There is one approach to hydrological similarity that has been used in studies of both runoff and evapotranspiration fluxes. This approach derives from the work of Kirkby (1975) and Beven and Kirkby (1979) who showed that under assumptions of a homogeneous and steady state recharge to the water table, local storage deficits can be scaled by the topographic index ($a/\tan\beta$), where a is the upslope area draining through a point per unit contour length and $\tan\beta$ is the local slope angle, used as a surrogate for the downslope hydraulic gradient. O'Loughlin (1981,1986) derived a similar "wetness index" independently.The distribution of the index could then be used to predict time variable saturated contributed areas and depths to the water table. All points in a catchment with the same value of the index are predicted as having hydrologically similar responses. This distribution function approach forms the basis of the hydrological model TOPMODEL , the history and background of which is described more fully in Beven et al. (1994).

Beven (1986) later showed that variability in soil characteristics could be incorporated into a combined soil topographic index, $\ln(a/T_0\tan\beta)$ where T_0 is the saturated transmissivity of the soil, while Quinn et al. (1994) showed how the predictions of variable depths to the water table could be used in spatially variable predictions of evapotranspiration fluxes. This approach also served as the basis for a theory of hydrological similarity of runoff production and flood frequency curves, involving 5 dimensionless variables, described in Sivapalan et al. (1987), Wood et al. (1990), and Larsen et al. (1994). In all cases the assumptions made in the derivation of the TOPMODEL theory mean that it will be most appropriate in areas where there are consistent downslope flows in a relatively shallow soil underlain by an impermeable bedrock.

Approaches to the Scale Problem

No adequate theory exists in hydrology for defining an appropriate description of hydrological processes at a given scale, nor for making use of information at one scale for a description at a smaller or larger scale. Most approaches to this problem, as noted above in the application of SVAT models, have involved taking a small scale theory and using it at a larger scale without consideration of the scaling effects involved except to assume that "effective" parameters can be used within that theory to reflect the heterogeneity of characteristics and response to be expected at larger scales. This "aggregation" approach to the scale problem has been criticised previously by Beven (1989) on the basis both that the small scale description may not be appropriate at the larger scale due to structural and preferential flow effects and that the nonlinearities inherent in the system mean that the concept of effective parameters is not generally tenable.

So, if such an aggregation approach is fraught with difficulties, is there a viable alternative "disaggregation" approach to the scale problem. A disaggregation approach implies a search for an appropriate description of the system at the scale of interest, be that point, hillslope, catchment or macroscale. The appropriate description should be expected to be itself scale dependent. By analogy with atmospheric models, this disaggregation description is essentially equivalent to an appropriate sub-grid scale parameterisation of the smaller scale processes; in terms of previous work in hydrological models, it is equivalent to an appropriate conceptual model of the processes at the catchment scale. Previous work on sub-grid scale parameterisations and conceptual models suggests that the disaggregation description must be parsimonious in its parameters and conditioned on the data available to determine values for the parameters (see for example Kirkby, 1975; Shuttleworth, 1988).

There is, of course, a plethora of catchment scale conceptual models in hydrology, most of which can be made to be reasonably successful in predicting discharges provided that some discharge data is available for calibration purposes. Some of these models are simply collections of linked storage elements that rely on the calibrated parameter values to reflect the heterogeneity of responses in the catchment. Others have tried to build in some consideration of the distribution of responses, either directly such as in TOPMODEL (see recently Quinn and Beven, 1993) and the distributed stores model of Moore and Clarke (1981, Moore 1985); or implicitly such as in the Xinanjiang/Amo model (Zhao, 1992; Dümenail and Todini, 1992). In addition, recent advances in SVAT parameterisations have started to use distribution functions of land surface parameters (eg Famiglietti and Wood, 1991; Avissar, 1992; Dolman, 1992; Koster and Suarez, 1992; Blyth et al., 1993).

It is worth noting here that the integration of a distribution function of responses to represent a larger area is a linear process, but that the nature of those responses and the model representing them may be nonlinear. In taking this approach, however, the representative distribution functions must reflect the critical sources of sub-grid scale variablility, which may vary with both environment and scale and may be identified either theoretically or empirically.

The Macroscale Parameterisation Problem

Consider the response of a large area. The "points" in that area have different characteristics of topographic position, soil and bedrock profiles, vegetation communities, and rainfall inputs. This variability of characteristics will be known only in very general terms, since only a limited amount of data will be available, particularly for soil, vegetation and bedrock characteristics. A model is required to predict the latent heat fluxes (or discharges) from that heterogeneous area. Assume, for the present, that it is accepted that a distribution function framework is the most promising approach for defining an appropriate sub-grid scale parameterisation. The full distribution function will be made up of the responses from every combination of point characteristics in the catchment, but , as noted above, many points may be effectively similar in their responses. It should not therefore be necessary to make computations for every point in the area. A much smaller number of "representative" calculations should be adequate to represent the range of point responses contributing to the areal response to some acceptable degree of accuracy.

The question then is to decide how to represent those point responses and which combinations of parameter values might be "representative". One approach to this problem is to take a similar approach to the aggregation strategy noted above, by trying to estimate representative parameter values for sub-areas or chosen points and hoping that these can act as effective values in reproducing the variability in the area as a whole. Such a strategy might not properly reflect the role of extreme values in the areal response since the tendency in chosing representative points is to select modal values. A variant on this strategy would be to estimate distribution functions for the joint occurrence of each individual parameter, for either the whole area or discrete sub-areas, and then integrate the resultant calculations over the multiparameter function. This statistical-dynamic approach has been taken, for example, by Entekhabi and Eagleson (1989). This approach can properly reflect the extremes of the sub-grid responses, provided it is based on an appropriate model, and that the data are available to estimate the parameter distribution function. Data availability is commonly a problem in this respect, particularly for subsurface parameters.

An alternative strategy is suggested by looking at the problem from a disaggregation viewpoint. What is the range of possible sub-grid responses to be expected in the area? What then is the minimal parameterisation to represent that range? How then can that parameterisation be calibrated given the available hydrological data (perhaps discharges in one or more subcatchments)? In principle, given enough computer time, it might be possible to run all possible parameter combinations for the sub-grid areal components or "patches" (be they points, hillslopes or hydrological response units), conditioned on some *a priori* choice of the range to be considered in each case. The result would be N simulated hydrological responses, summarised as predicted evapotranspiration and discharge records. Many of those N responses will be similar, either analytically from the structure of the model (such as through the combination of a, $\tan\beta$ and T_0 in the TOPMODEL structure) or practically in that they give very similar results. What is needed in the macroscale parameterisation is then to choose a combination or distribution of these N runs that consistently reproduces the macroscale response in so far as that can be verified by the available data. This approach is considered, using one particular patch model definition in what follows, although it is clear that it could also be extended to consideration of multiple model definitions.

A Representative Patch Model

A patch is treated as any area of land within a basin or GCM grid square that has predominantly similar properties. The definition of similarity in soil, vegetation, or topographic characteristics or a combination of these is still open to debate. The size and shape of a patch will be variable. The extremes of the area would be a patch equal to the size of the catchment or a patch equal to the size of a GIS pixel. It is envisaged that a macroscale SVAT model would be a series of patches that capture the dominant heterogeneity of the landscape under analysis. The number of patches should be as few as possible, but a maximum could well be a distribution function form already used within the TOPMODEL structure (which is usually 15-30 increments).

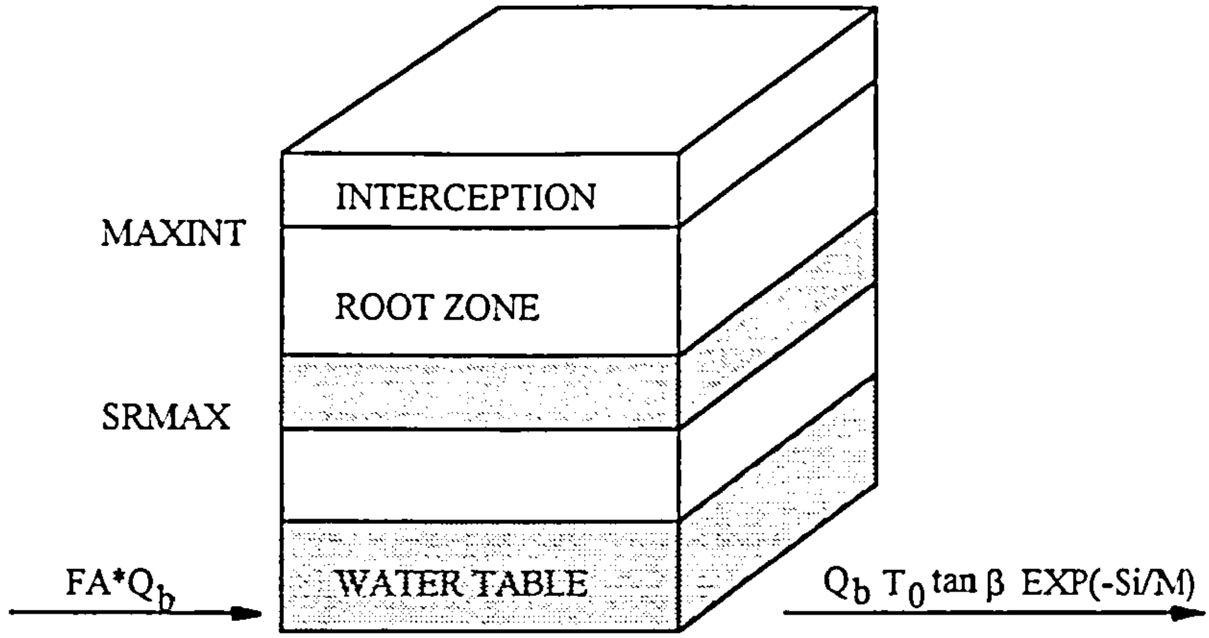
Figure 1 shows one conceptualisation of a patch model, containing a simple representation of the dominant hydrological processes that influence evaporation and drainage. In this form it is similar to some of the bucket type SVAT models used widely in the past but with one critical difference: that water will enter the patch by subsurface drainage from upslope and that water is also lost from the patch by downslope drainage. This input and output is necessary as the patch will have its own position in the landscape. For example, if the patch were a riparian area then it is expected that the water table will be recharged by flow from upslope while close to the divide, such a contribution to the water balance of the patch will be minimal. The control of the position of the water table in this may have an important effect on both evapotranspiration and runoff as discussed in Quinn and Beven (1993) and Quinn et al.(1994).

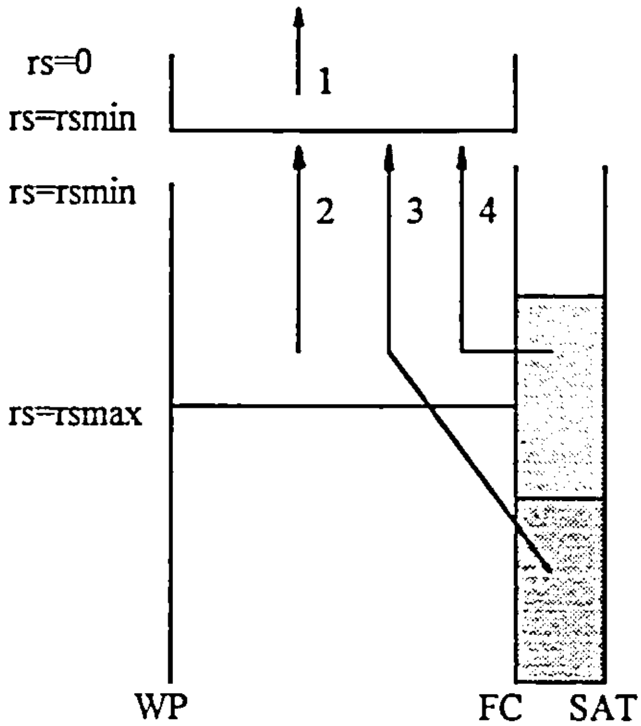
In what follows, we simplify the treatment of the downslope fluxes by following a similar set of assumptions to those used by TOPMODEL (Beven and Kirkby, 1979; Quinn and Beven, 1993; Beven et al. 1994) in which downslope fluxes are assumed to be in equilibrium with a recharge rate calculated from total hillslope discharge averaged over some area, lateral transmissivity is assumed to be an exponential relationship between subsurface storage or storage deficit, and the downslope hydraulic gradient is assumed constant. With these assumptions, it follows that the outflow per unit contour length is given by:

$$Q_b = \tan\beta \, T_o \, \exp(-S_i/m)$$

where $\tan\beta$ is the hydraulic gradient (in TOPMODEL approximated by the ground surface slope but here only assumed a constant), T_o is the transmissivity of the soil when saturated and S_i is a storage deficit due to drainage in the patch. The parameter m is a

THE PATCH MODEL.





1= Evaporation from the interception store,
2= Evaporation from the root zone store,
3= Evaporation from the water table when the water table is in the root zone.
4= Evaporation from the water table by capillary rise (Eagleson, 1978)

scaling parameter for transmissivity for which a catchment average value can be derived from a recession curve analysis. In the analysis that follows m has been fixed in this way. In addition, T_0 and $\tan\beta$ only occur in the analysis as a product so that they have been combined as a single variable, thereby further reducing the number of parameters to be considered.

Keeping the same assumption that subsurface flow upslope of the patch is always in a quasi-steady equilibrium with the current drainage, lateral drainage from upslope can be calculated for a unit pixel size as :

$$Q_{in} = Q_b \, FA$$

where *FA* is the ratio of the area per unit contour length draining into the patch, to that draining out of the patch. For a unit square patch, with an upslope contributing area of *a*, then $FA = a/(a+1)$. For a convergent patch, *FA* could be greater than 1. The net drainage flux will be of great importance to the maintenance of the water table in the patch; this is the significant difference between this and most other SVAT models, although TOPMODEL derived landscape scale water balance models have also been described by Band et al., 1991; Famiglietti and Wood, 1991; and Famiglietti et al. 1992). Note that with an application of TOPMODEL the a values are normally derived by topographic analysis under the assumption that the upslope contributing area extends to the divide. That assumption is not necessary here, the *a* value need represent only some **effective** upslope contributing area. It remains, however, necessary to assume that this is constant over time.

The evapotranspiration calculation is based on the Penman-Monteith equation with a constant aerodynamic resistance coefficient according to the vegetation structure of the patch. When the interception store is full to its capacity *MAXINT* the canopy resistance (r_s) is put to 0 s/m. Canopy resistance is assumed to increase as the interception store to a value *rsmin* (which is the value for evapotranspiration from a well watered soil with a dry canopy). The value of *rsmin* (note fixed parameters are put in lower case while parameters that vary are put in upper case) has been fixed as 50 s/m which is a typical value. When the interception store is full, moisture will overflow to fill the root zone which has a maximum capacity of *SRMAX*. The canopy resistance is increased linearly as the root zone dries to a value of *RSMAX* which will also depend on the vegetation type.

Calculated evapotranspiration from the root zone depends on the current canopy resistance. However, the total evaporation from the patch may be higher than this if water is available from the water table, due to the water table being within the root zone or by capillary rise, up to a maximum of the potential dry canopy rate.

The amount of water available from capillary rise is controlled by the formula of Eagleson (1978), where the potential upward flow is calculated for a particular soil. Four soil types, representing a range from clay to sand, have been used here with hydraulic properties taken from Bras (1990). Water in excess of the capacity of the root zone is routed to the water table using a simple time delay per unit deficit approach as used previously in TOPMODEL (Beven et al., 1994; parameter *VTD* in Table 1).

One remaining parameter is the *REFLEV* (reference level) parameter which controls the reference depth for the exponential transmissivity function. This concept, first used in Quinn et al. (1991) allows for a water table that is relatively deep due to local conditions. The REFLEV displacement can used to model a groundwater situation without using a complex groundwater model. Small values of REFLEV increase the likely contact of the water table with the root zone whilst deeper values reduce the contact of the water table with the root zone and alter the time delay calculation in the unsaturated zone.

Model dynamics with Environmental Forcing

Figure 2 shows a set of three runs using environmental variable data taken from the Réal Collabrier catchment, Southern France. In these initial runs a period of 1000 hours (2 months) has been used with a sequence of rainstorms during a period of high potential evapotranspiration. Lacking other data, the specific humidity deficit and the aerodynamic resistance have been set to reasonable constant values. Three example simulations are shown in Figure 2 to demonstrate the range of operation of the model. The predicted evapotranspiration plot shows three components of evapotranspiration for each of the three simulations. The interception losses during and immediately after each storm are quite clearly seen. The losses from the root zone (two of which have been highlighted in heavy black lines) show the gradual decline as the root zone dries, with a diurnal variation due to the daily pattern of radiation inputs. Of the three scenarios only one shows significant fluxes from the water table (highlighted as a dark dash-dotted line). If the water table is shallow and the root zone begins to dry then evapotranspiration fluxes are being maintained by upward flux from the water table.

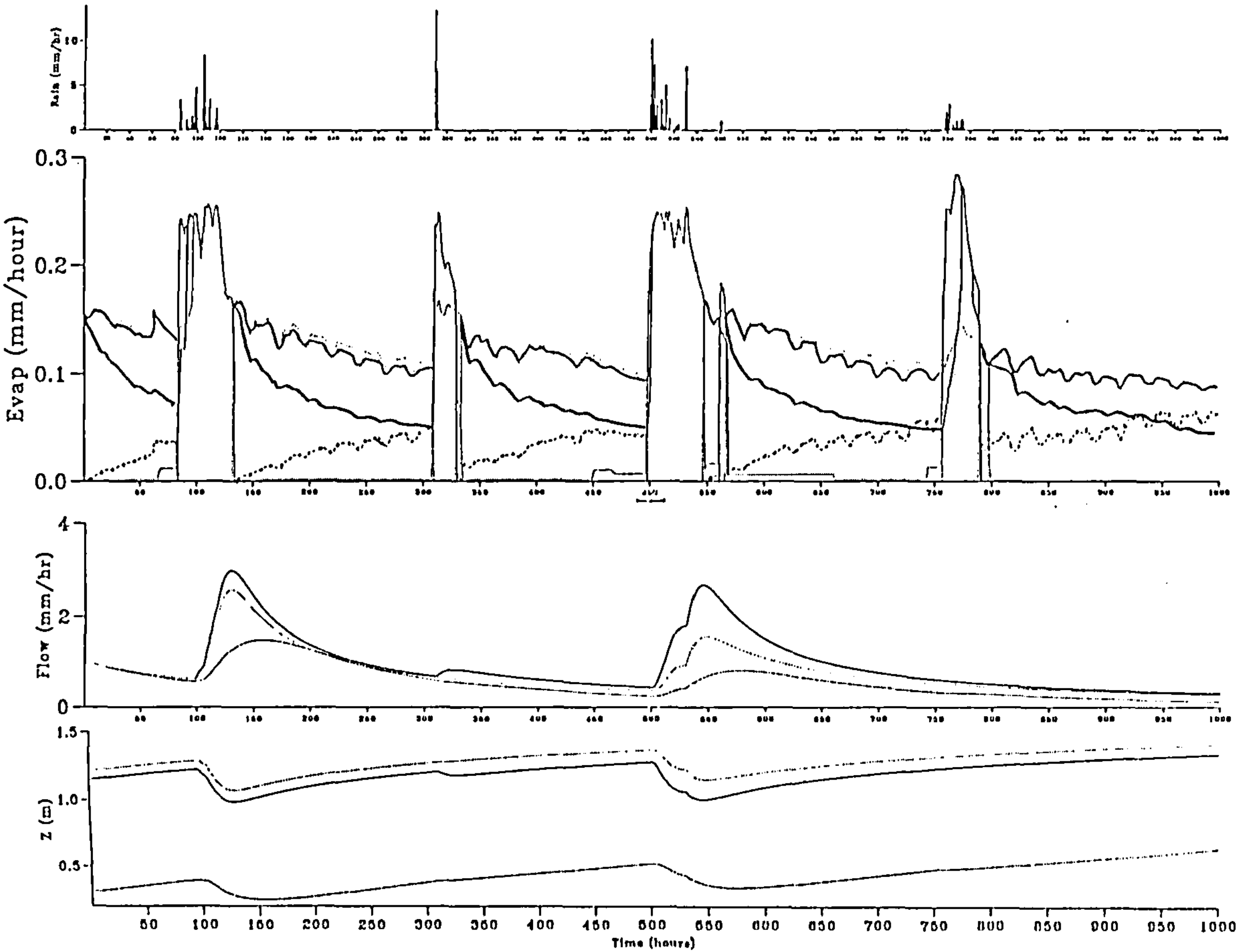


Figure 2. The operation of the patch model against time for 1000 hours of data taken from the Réal Collabrier catchment, Southern France using three diferent sets of parameter values.

The plot of calculated subsurface flow from the patch shows the sensitivity of the model to the recharge of water to the water table and the effects of the unsaturated zone time delay. The lower plot shows how the water table depth (z) varies in the three simulations. All the predicted water tables have a similar pattern but the higher value of the *REFLEV* parameter in two of the cases results in a displacement of the water table to a different range of depths.

Monte Carlo Simulations

Given the form of patch model described above the range of possible behaviours has been explored using Monte Carlo simulation, taking random sets of 8 parameters (including the soil type) from uniform distributions across chosen ranges, as set out in Table 1. The same single environmental forcing record as used for Figure 2 was used in all 10000 Monte Carlo realisations. In each realisation the root zone was taken as initially full, but the initial subsurface discharge $Q_b(t=0)$ was set a value of 1 mm/hr for all simulations, as if this was an observed value at the catchment scale.

Table 1. Patch model parameters varied in the Monte Carlo experiments

Parameter	Lower Limit	Upper Limit
RSMAX (s/m)	300	600
SRMAX (m)	0.02	0.1
MAXINT (m)	0.001	0.01
FA	0.1	1.0
REFLEV (m)	0.01	1.0
$T_o tan\beta$ (m^2/hr)	0.005	0.02
VTD (hr/m)	0.001	1.0
SOIL	1 (clay)	4 (sand)

Table 2 shows the ranges of each evapotranspiration component generated from the model. It can be seen that the total evapotranspiration, even with a single set of environmental forcing data can vary by a factor of 2. Figure 3 shows how the total evapotranspiration varies with parameter value for four of the parameters in the model. Each point represents a single realisation of randomly chosen parameter values. The extreme values and wide scatter of the points reflect the efffects of parameter interaction on the total calculated evapotranspiration.

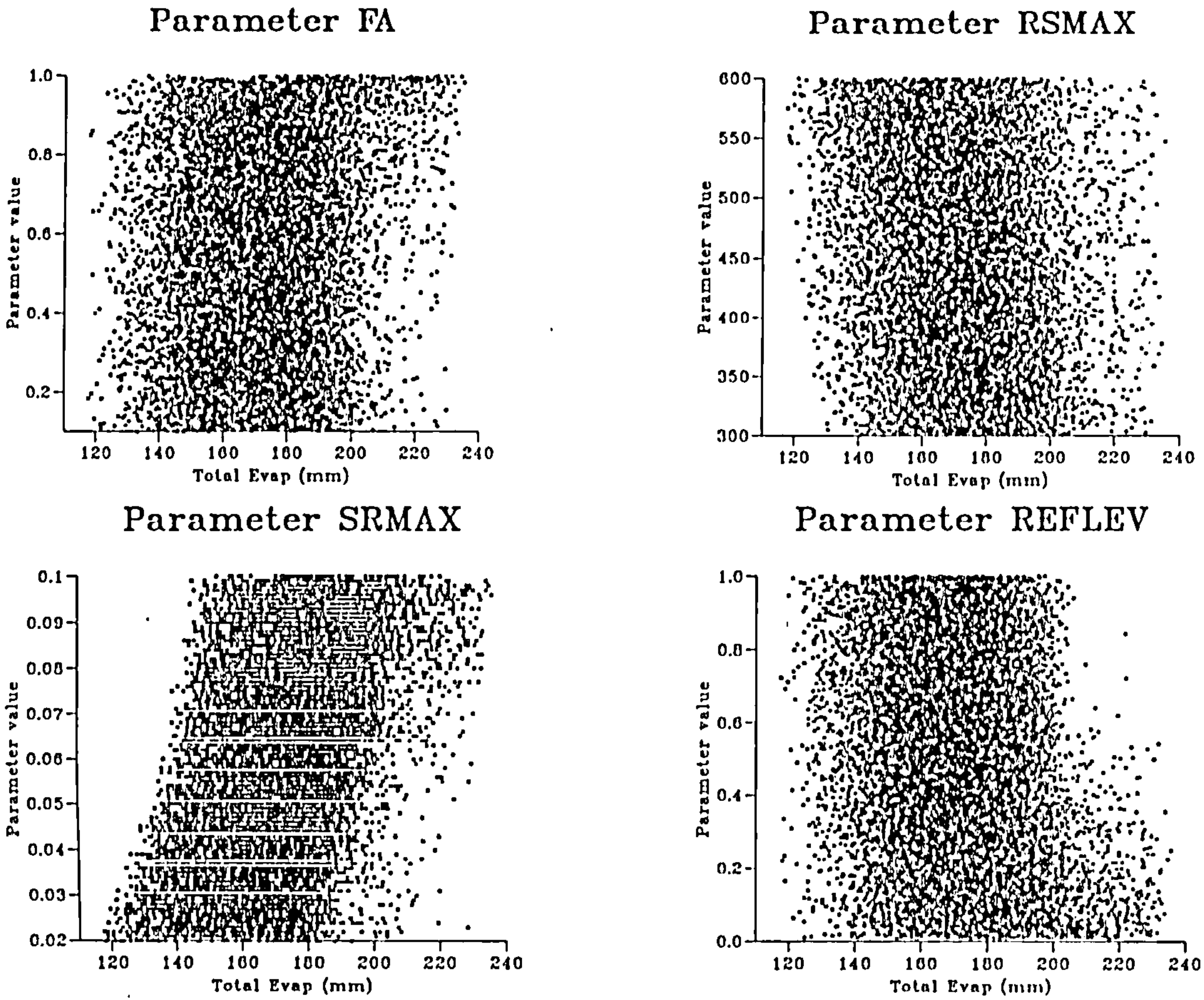


Figure 3. The variation of total evapotranspiration with 4 of the model parameters, demonstrating the effects of parameter interaction. Results from 10000 Monte Carlo simulations.

Table 2. Evaporative dynamics of the patch model given 8 varying parameters, run for 1000 hours, based on 10000 separate simulations

	Average	Maximum	Minimum
Total evapotranspiration (mm)	171.5	235.2	117.6
Root zone evapotranspiration (mm)	81.7	109.7	53.1
Interception store evaporation (mm)	85.9	127.7	45.9
Evapotranspiration supplied by water table (mm)	2.4	51.2	0.0
Evapotranspiration supplied by capillary rise (mm)	1.4	64.8	0.0

The behaviour of even this simple model is clearly very complex but it is clear that if the interest is in predicting the range of evaporative fluxes within the landscape it should be possible to reduce the dimensionality of the system. An examination of this possibility has been made by dividing the observed total evapotranspiration into 10 subsets. Each realisation can then be assigned to one of these subsets, each of which will contain a different number of realisations (in this case the numbers range from 79 to 2103 for the 10000 simulations). The distribution of parameters in each subset has then been plotted to examine the sensitivity of total evapotranspiration to that parameter. Figure 4 shows the results, where subset 1 is for the lowest range of total evapotranspiration (117-129 mm) and subset 10 the highest (223-235 mm). The normalised X axis in each case represents the cumulative frequency for the parameter in each subset. A straight line reflects parameter insensitivity for that evapotranspiration range. A marked departure from the straight line indicates a zone where a parameter is sensitive. For the four parameters shown the following comments may be made:

1. *FA*, the fractional area recharging the patch. This shows that for most of the evaporative range the parameter is insensitive. But for the higher evapotranspiration subsets, the prediction becomes sensitive to high values of *FA* which will help to maintain the water table during dry periods.

2. *RSMAX*, the maximum canopy resistance. This shows less sensitivity but for the lower evapotranspiration subsets there is some response to higher values of *RSMAX*.

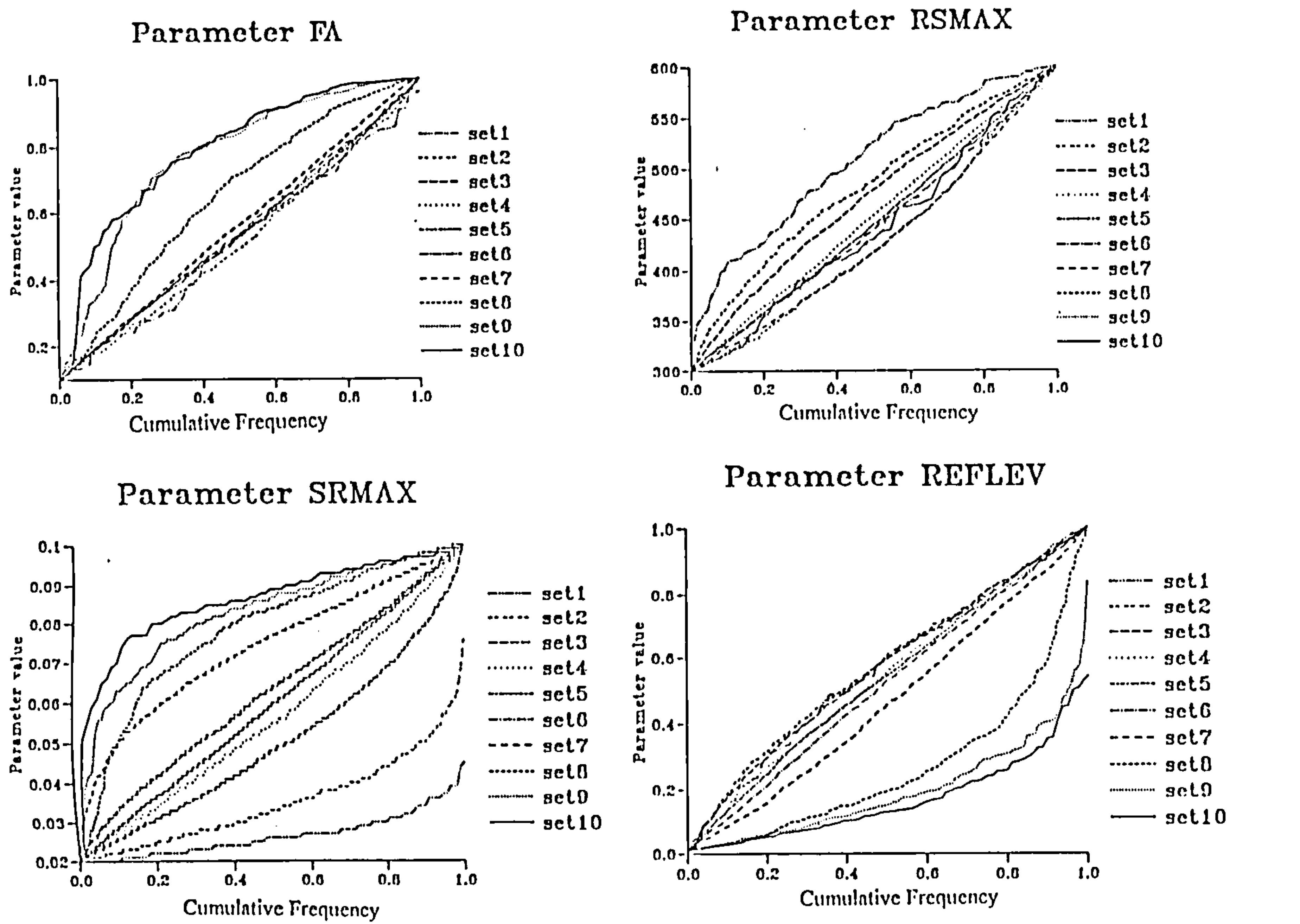


Figure 4. Cumulative distributions of parameter values for each subset class of cumulative total evapotranspiration. Subset 1 is the lowest range (117-129 mm), subset 10 the largest (223-235mm).

3. *SRMAX*, the capacity of the root zone store. This parameter shows the greatest parameter sensitivity across the complete range of evapotranspiration subsets. As would be expected, the low *SRMAX* values are associated with low evapotranspiration totals and high *SRMAX* values strongly influence high evapotranspiration (but note still the spread of values for the *SRMAX* plot in Figure 3).

4. *REFLEV*, the control depth for the water table. For most of the evapotranspiration range the *REFLEV* parameter is insensitive. However, as the higher totals are approached the parameter becomes quite sensitive especially to lower values of *REFLEV*, which increase the likelihood of support for evapotranspiration from the water table.

This type of analysis allows us to describe under what conditions any parameter is sensitive or redundant. For this environment the root zone size is the most dominant parameter (as it is generally for single lumped bucket models). However giving the drying potential of the environmental forcing data it is seen that there is also some sensitivity to parameters influencing the water table. This type of analysis can be used to suggest ways in which effective or "index" parameter sets might be used to represent the range of behaviours needed in any given environment.

Discussion

The results presented here have concentrated on functional similarity in the prediction of the evapotranspiration components of the water balance. It has been shown that multiple parameter sets result in similar functional behaviour suggesting that, in a disaggregation approach to sub-grid scale parameterisation problem, it might be possible to reduce the dimensionality of the parameter identification problem for predicting catchment or landscape scale fluxes. One possibility for doing this would be by choice of suitable sets of "index" parameters. This is, however, only one component of the problem of estimating landscape fluxes. Even if it is accepted that the aggregated flux can be treated as a linear sum of the index parameter set fluxes, ie. treating the component fluxes as independent tiles in a multiple patch model, it will still be necessary to estimate the weighting coefficients to apply to each component to represent the frequency of occurrence within the area under consideration.

Ideally these coefficients would be defined by the pattern of vegetation, soil, topographic and geologic characteristics within the landscape. In practice there may be considerable uncertainty in the estimation of the frequency of certain parameters, even with this very simple parameterisation (eg. *REFLEV*, T_o, effective *FA* values) due to problems of both heterogeneity and observability. However, in principle at least, the disaggregation parameterisation problem might be reduced to the problem of estimating a set of N linear weighting coefficients, where N might be a relatively small number. For evapotranspiration estimates alone, in some environment where the surface is either consistently wet or consistently dry N may be equal to 1. Where there is a valley bottom or water body surrounded by a dry landscape, N = 2 may be sufficient. In many environments, such as the Mediterranean climate used here it may be necessary to have a range of index parameter sets to represent adequately the variability of the fluxes in the landscape. In this disaggregation approach, both the nature of the patches or index parameter sets considered and the set of associated weighting coefficients must be expected to be scale dependent.

This preliminary study has been greatly simplified in that it has not considered spatial variability in the imposed environmental forcing conditions, nor other predicted variables such as discharge. Such variability might be expected to increase the number of index parameter sets required in the patch model, especially if the timing of discharge predictions is important. However, the important simplifying principle of mapping the variability onto a range of functional behaviours, with appropriate weighting coefficients should remain. While some of that mapping might result from GIS overlays, other aspects will almost certainly require stochastic components, of which more later. This study has been a first attempt to show which variables may be the most significant in the nature of that mapping.

Acknowledgments

This work has been supported by the UK NERC Terrestrial Initiative in Global Environmental Research (TIGER) Programme. Jacques Lavabre and Flavie Cernesson of CEMAGREF, Aix-en-Provence and James Fisher of IEBS, Lancaster are thanked for the provision and preparation of the RÈal Collobrier data as part of the European Community DM2E project.

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GLENAMOY AND CLARA BOG RESEARCH

ASPECTS OF BLANKET BOG RESEARCH AT GLENAMOY

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Abstract

The Glenamoy project was established to investigate the reclamation of deep blanket peat for agriculture. At the time, the emphasis was on increasing production and there was not much consideration given to environmental issues. The research programme was concerned with crop and animal production. The crop programme was concerned mainly with fertiliser requirements. The programme on animals dealt mostly with the nutrition and management of sheep and cattle. A separate drainage investigation was also undertaken. The present report deals with drainage and related matters. Much information was obtained but the main thrust of the project was abandoned after a few years because of a severe epidemic of liver fluke in sheep, and because it was realised that both animal and machine traffic caused serious problems on the soft peat in the wet climate at Glenamoy.

Introduction

Forty years ago when Glenamoy first came into prominence the interest was all on increasing production, and environmental considerations were scarcely heard of. At that time the Department of Agriculture had a slogan "One more cow, one more sow, one more acre under the plough". So what started the Glenamoy project?

After the war the Irish Sugar Company undertook an operation to produce grass meal on a raised bog at Gowla in Co. Galway. This was initially successful and it soon became a political matter. The perceived wisdom was that here was a previously useless bog now producing a very useful product. There are more than three million acres of unused peat in the country; why not cash in on this potential richness? More specifically there was one quarter of a million acres of unused peat in Erris in north Mayo, an area of high unemployment and poor resources. At that time the party in power established a company, Min Fheir Tta, to produce grass meal on bog in Erris. The opposition shadow Minister for Agriculture opposed this move on the grounds that enough was not known about producing grass meal on blanket bog to justify the Company and he promised that on the change of Government he would establish a research station in Erris to get the necessary information. The blanket peat at

Glenamoy was completely different from the raised peat at Gowla. After the next election there was a change of Government and the new Minister for Agriculture established the Peatland Research Station at Glenamoy.

The Research Programme

The general objectives of the station were fairly broad and included, the growing of various crops, in particular grass, the production of sheep and cattle and the investigation of the chemical and mineral requirements of crops and animals. From the beginning there were serious problems with trafficability both animal and machine and these will be referred to later. While most of the common farm crops were grown and their fertilizer requirements, including trace elements were established, it soon became evident that grass was the only crop that was suitable to the raw blanket peat. A farm of 140 acres was established to evaluate the feasibility of general farming on blanket peat and a programme on drainage was undertaken. While all aspects of the research programme were pursued vigorously and were very successful in terms of providing information I will concentrate here on the drainage programme and its results. At the outset three major objectives were identified:

1. To determine the drain depth and spacing required for the crops grown at Glenamoy.
2. To devise methods of achieving the required drainage.
3. To determine the effects of this drainage on the peat in relation to crop requirements. Water balance studies arose later as an addition to the general drainage investigations and while some information on water balance was obtained it was not regarded as a principal component of the drainage study.

Results

Before discussing the results obtained it is first necessary to consider the weather at Glenamoy, more specifically precipitation, and also the main physical properties of the Glenamoy peat. Annual precipitation at Glenamoy is about 1200mm and it has ranged from less than 1100mm to more than 1500mm. Rain falls on about 250 to 270 days per year. The wettest period is generally between October and January and the driest between February and July, but there is no definite dry period. The most important properties of the peat in relation to drainage are hydraulic conductivity and the inherent moisture characteristics of the peat. Field measurements show the saturated hydraulic conductivity and the inherent moisture characteristics of the to be in a range between $1x10^{-5}$ and $1x10^{-6}$mm/day. Laboratory measurements by Galvin and Hanrahan (1967) showed a range of $1.2x10^{-6}$mm/day at a tension of 0.3m water to $3.5x10^{-7}$ mm at a tension of 1.2m water. In effect draining the peat reduces its hydraulic conductivity. The volumetric moisture content of the undisturbed peat ranges from 90 per cent at the surface to 93 to 95 per cent at greater depths.

Drainage requirements: The above parameters indicated that the best method of determining drainage requirements was by direct experimentation. A simple field experiment based on various drain spacings and depths combined with watertable measurements and crop performance showed that a spacing of about 2.5m at a depth of about 0.8m was necessary and adequate for the crops grown at Glenamoy. Drainage in relation to trafficability will be referred to later on.

How to achieve the drainage: The main method of drainage used in peat for commercial purposes was to plough open drains. Such drains were about 0.75m wide at the surface and at a spacing of 2.5m they would reduce the workable surface by about 30 per cent, as well as restricting machines. It was therefore necessary to provide some form of covered drain. Conventional tile drains were too expensive and there were also problems with stability in the soft peat. Several methods of drainage were tried and eventually two types of drain were selected: Tunnel drains and Gravel on polythene drains.

Tunnel Drains: A sketch of the tunnel plough is shown in Figure 1 and the method of operation in Figure 2. The plough is of simple construction. The breast plate splits the peat like an ordinary plough. The lower part of the plough consists of an orifice which connects to the bog surface through a curved duct. The forward motion of the plough forces a rectangular section of peat into the orifice and through it to the surface. In operation a ribbon of peat is excavated and, as the plough passes the top layer that has been pushed aside, returns to its original position leaving a small tunnel in the peat. It is from this feature that the name "tunnel plough" derived. A potential major drawback with this plough derives from the variability in the peat. Where the peat is too firm it does not move freely through the duct and the plough clogs. Where peat is too soft the drains are unstable and close in very quickly.

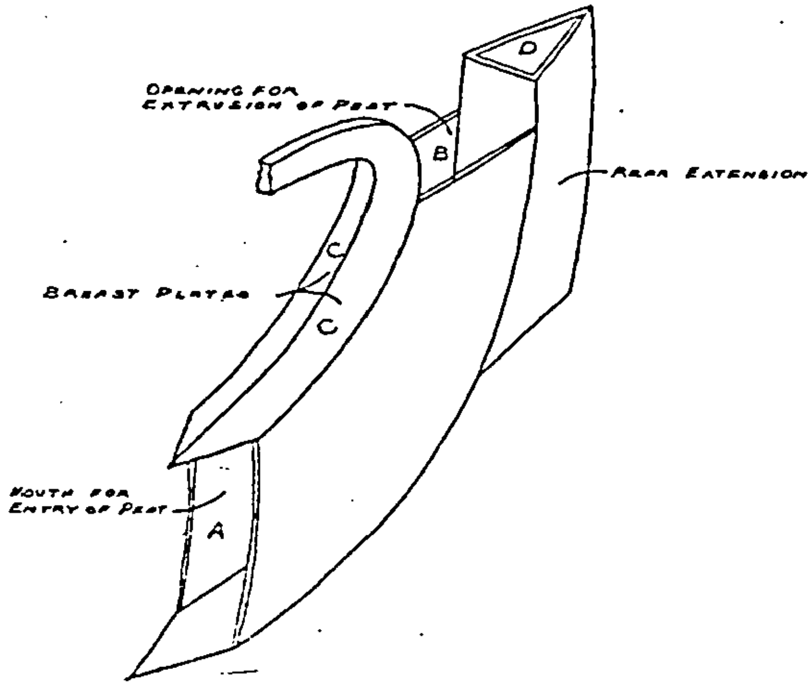

Figure 1 The tunnel plough

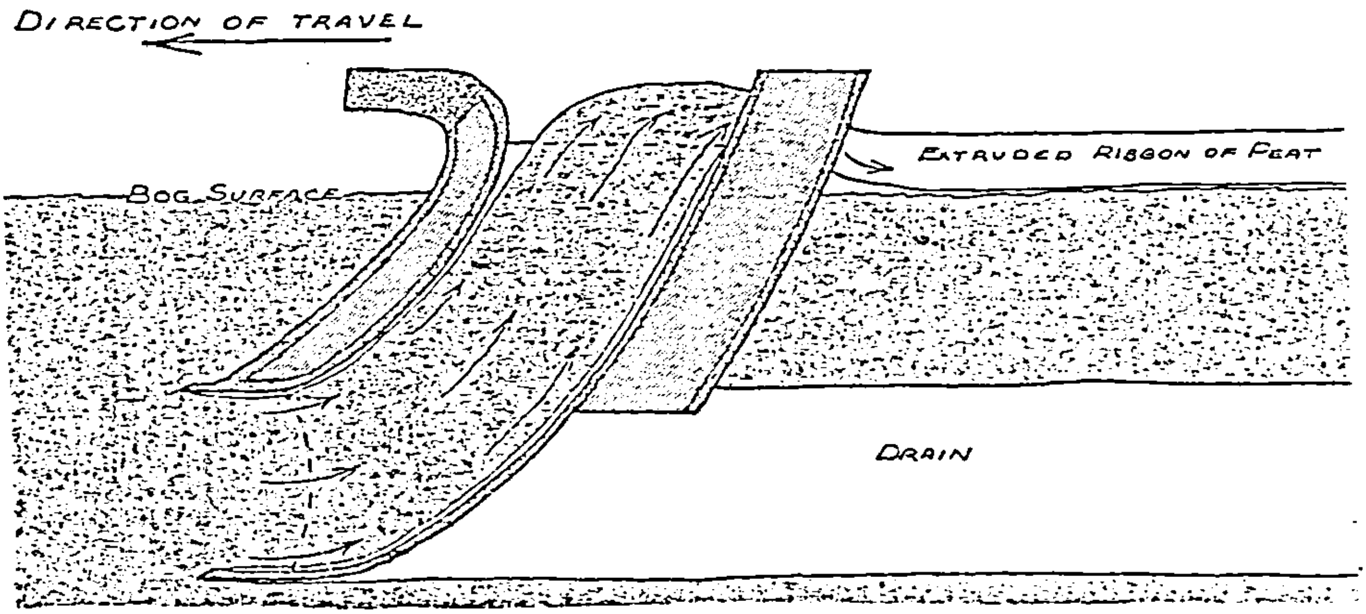

Figure 2 Method of operation of tunnel plough

A simple test was devised to assess the suitability of peat for tunnelling. A Pilcon hand vane was modified by enlarging the flanges on the instrument - Figure 3. By taking readings in a wide range of peats of different consistency and examining plough performance across the range, a scale was devised which predicted the suitability or otherwise of the peat for tunnelling - Figure 4. The cross-section of the plough duct and orifice were modified to arrive at practical dimensions. Other modifications were also indicated but were not possible for reasons that will be considered later. The principal modifications were: the replacement of the base of the duct with a moving or possibly a driven belt to reduce internal friction; making the duct cross-section slightly larger than the orifice; also to reduce friction; and finally to devise a suitable mounting method to give a better distribution of weight. With the 3-point linkage used the downward pressure of the peat in the plough transferred the centre of gravity of the system towards the rear and made the operation very inefficient. Many other minor modifications, not listed here, were also indicated.

Gravel on polythene drains: The gravel on polythene drain was developed to overcome some of the problems of tunnel ploughing, particularly lack of stability in soft peat. A diagram of the drain is shown in Figure 5, and of the prototype plough that installed it in Figure 6. The drain consisted of a section of gravel supported on a strip of polythene. Consideration was given to determining the optimum size of gravel and the drain cross-section. The cross-section is related to length of drain and gradient of the peat and the hydraulic conductivity of the gravel and these aspects are easily determined. As in the case of the tunnel plough several modifications were required but were not made. The main requirements related to proper mounting and the transport and provision of gravel on a continual basis on the soft peat. Despite these difficulties about 60 acres was drained by this method - quite successfully within the limitations of Glenamoy.

After a few years of research at Glenamoy a very serious problem arose with Liver Fluke in sheep. The peat, modified by lime, fertilizers and other minerals in the wet environment of Glenamoy provided an ideal environment for the small snail that is the secondary host of the Liver Fluke. Natural predators of this snail were absent in the early years at Glenamoy. There was an explosion in the snail population and correspondingly the fluke population reached near epidemic proportions. There was

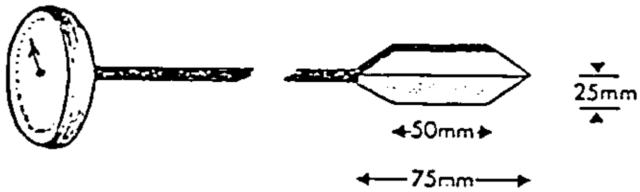

Figure 3 Modified vane

Vane Reading	Comment
65	
90	No sod
90	
80	Broken sod
185	Bog firm - good tunnel formed
90	Broken sod - interrupted tunnel
135	Broken sod - interrupted tunnel
95	
110	
100	No tunnel
125	
140	
140	Interrupted tunnel
130	
140	
250 to 290	Very good tunnel formed over 100 m +

Direction of travel

Total length of run was about 300m. Vane readings were taken at varying intervals depending on quality of sod extruded.

Figure 4 Vane readings and suitability for tunnelling

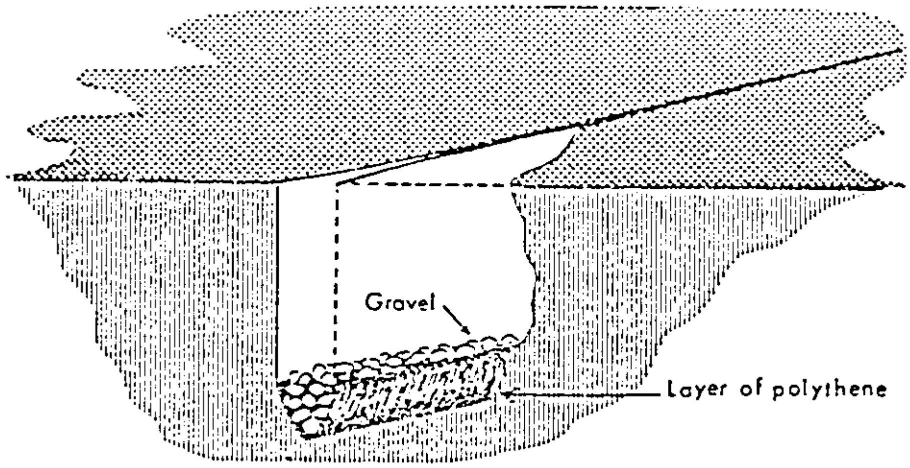

Figure 5 The gravel drain

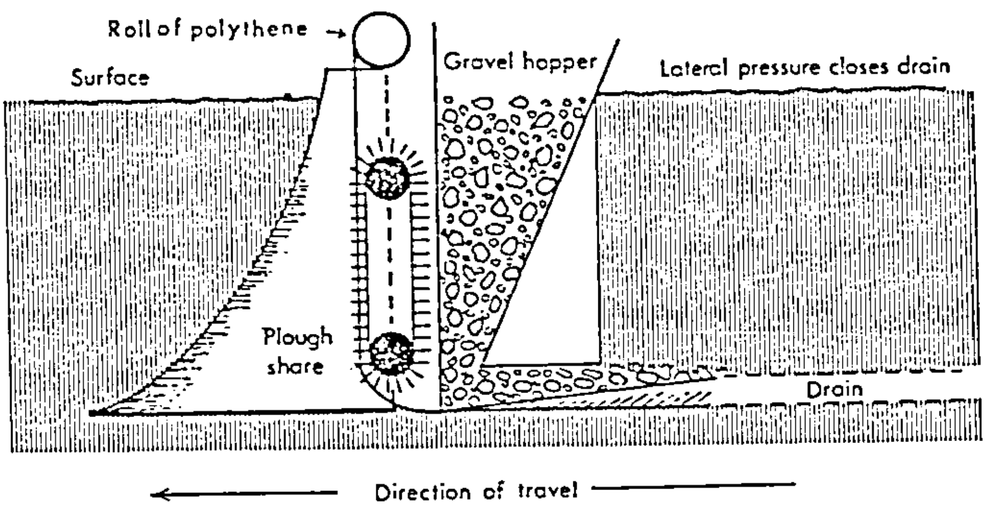

Figure 6 The gravel drain plough

also a growing realisation that while most common crops could be grown, their utilisation by animals or by mechanical means was very difficult on the soft peat. As a result, interest in the Glenamoy experiment waned rapidly. This is the principle reason why the ploughs were not developed more fully. Good progress was subsequently made in fluke control but by then the major impetus of the Glenamoy programme had been abandoned and it was not recovered.

Some Physical Properties of The Peat

Water balance studies commenced at an early stage. At the outset they were intended solely to provide background information to the drainage studies. They were confined to two fairly small areas, a drained and an undrained one and initially were designed to examine outflow patterns on raw peat and under grass.

Effects of drainage and utilisation on the peat

When considering the water relationships of peat it is necessary to examine the effects of drainage and methods of utilisation on the peat itself. At Glenamoy it was apparent that differences were arising and samples were collected from four different sites as follows:

(a) Undrained and unutilised peat

(b) Forestry - open drains approx 0.3m deep

(c) Forestry - Tunnel drains approx 0.6m deep

(d) Grazed pasture - various drains approx 0.6m deep

Water content was determined at the various sampling depths, at the time of sampling and after saturation in the laboratory. Air content was determined at sampling, and at various moisture tensions. Bulk density and hydraulic conductivity were also determined on peat cores. Some selected data are given in Tables 1, 2 and 3.

In Table 1 water content at sampling and at saturation in the laboratory, and air content at sampling and at two moisture tensions in the laboratory are given for two depths for four drainage and utilisation conditions.

Table 1: Air - Water relationships of peat samples

Sampling depth (m)	Water content (% vol) Sampling	Water content (% vol) Saturation	Air content (% vol) Sampling	Air content (% vol) 0.2m (tension H_2O)	Air content (% vol) 0.6m (tension H_2O)
		A - Undrained, Unused			
0.15m	86.5	93.6	7.0	10.7	19.9
0.6m	92.7	94.7	2.0	14.2	29.7
		B - Forestry - Open drains 0.3m deep			
0.15m	79.9	92.6	12.6	12.1	16.3
0.6m	91.5	94.0	2.5	8.0	20.9
		C - Forestry - Tunnel drained 0.6m			
0.15m	36.7	92.2	55.6	23.3	31.6
0.6m	75.1	92.2	17.1	10.0	19.8
		D - Grazed pasture - Drained at 0.6m			
0.15m	80.7	91.4	10.7	5.3	6.9
0.6m	86.8	91.5	4.7	5.2	7.3

In Table 2 dry bulk density values are given for samples from the same locations as those in Table 1.

Table 2 Bulk density of peat samples - (Kg/m^3)

		Forestry		Grazed
Sample depth (m)	Undrained and Unused	Open drains 0.3m	Tunnel drains 0.6m	Drained 0.6m
0.15	85	97	101	112
0.6	69	78	102	110

In Table 3 hydraulic conductivities of undisturbed cores are given for samples from the same location as those in Table 1.

Table 3 - Hydraulic conductivity of peat cores - mm/day

	Undrained	Forestry		Grazed
Sample Depth m	unused	Open drains 0.3m	Tunnel drains 0.6m	Drained 0.6m
0.15	35	8	1.7 X 10^4	29
0.6	22	47	1.2	1

The following are the major points to emerge from the tables :

Moisture content: While the moisture content is similar in all samples at saturation, it is much lower in the tunnel drained one at sampling than in the others, although there are differences between all samples **(Table 1)**.

Air content: A similar trend is obvious in air content. The tunnel drained area is well aerated, while the grazed area is less well aerated than the other sites **(Table 1)**.

Bulk density: There is an increase in bulk density from undrained through forestry to pasture **(Table 1)**.

Hydraulic conductivity: While hydraulic conductivity in the undisturbed peat is higher than average for the general area, the main interest is in the changes that have taken place with drainage and utilisation. There is a vast increase at 0.15m in the tunnel drained forest and a decrease at 0.6m in the same site as well as in the grassed area **(Table 2)**.

When all factors are taken together it is suggested that the following effects have occurred **(Table 3)**:

(a) **Well drained (tunnelled) forest:** In this site there was a very prolific root development to about 0.6m deep, no doubt as a result of improved aeration, and it appears that the root matrix held the peat in place thus limiting subsidence so that much of the shrinkage took place in situ, thus increasing the volume of large poles. This reduced the overall moisture content and increased the hydraulic conductivity, while at the same time there was sufficient general settling to increase the bulk density.

(b) **Grazed pasture :** Drainage was also effective on this site but the roots were not as prolific as in the forestry. A secondary factor was the grazing animals' hooves. As dewatering took place, and especially in dry weather, the animals hooves compressed the peat, expelling air from the pores, thus reducing overall porosity and increasing bulk density. The very low hydraulic conductivity at 0.6m in the two well drained sites is attributed to some dewatering of the peat at this depth with compaction. The consequence of these factors on utilisation was evident on the site. Under forestry, when the peat is drained, aeration is improved and the effect is permanent. Under grazing there is some improvement in peat strength in dry weather but as soon as rain

returns the peat surface re-wets rapidly with corresponding problems of poaching and rush infestation. Because the Glenamoy blanket peat is highly decomposed and has very few long fibres even moderate re-wetting results in a
rapid and serious loss of strength. With the frequent rainfall at Glenamoy even well drained grassland is difficult to manage and is unsuited to heavier animals. In fact sheep are the indicated animals but they are very subject to fluke infestation. It also poses severe problems for machinery operation because the wet peat is weak and is easily damaged by conventional farm machinery.

General conclusions : The Glenamoy programme provided the necessary information for the production of crops and animals on blanket peat, and also the yield potentials. It also showed the limitations of farming in such conditions. The most serious were animal parasites and trafficability.

The Water Balance

Water balance studies commenced at an early stage. At the outset they were intended solely to provide background information to the drainage studies. Because they were not specifically designed to determine the water balance, some of the investigations were not as detailed as a precise water balance study would have required. However, they provide general information on the water balance at Glenamoy. They were confined initially to two small areas of peat, 0.35 ha each, one drained and one not drained. The objective was to compare outflow from the drained area with surface flow from the undrained area and thus determine any differences in quantity and pattern between the two areas. Initially the areas under study were under the natural bog vegetation, but subsequently grass was sown with no measured effect on the results. At a later stage a similar study was undertaken in a forested area and here the results were more dramatic.

THE WATER BALANCE ON NATIVE VEGETATION AND GRASS

The main results from the study are summarised here. Table 4 gives a summary of annual values of precipitation, measured potential evapotranspiration, drain flow and surface run-off.

Table 4 Summary of yearly water balance 1969-1973 inclusive

Year	Ppt(mm)	P.E.	Drain flow(mm)	Surface flow(mm)	Water balance	
	(a)	(b)	(c)	(d)	a-(b+c)	a-(b+d)
1969	1231	489	962	742	-220	0
1970	1436	454	1126	784	-144	193
1971	1117	556	749	471	-188	90
1972	1181	500	804	496	-123	185
1973	1352	473	971	566	-92	313

Table 5 Summary of Summer water balance (May-Oct inclusive)

Year	Ppt(mm)	P.E.	Drain flow(mm)	Surface flow(mm)	Water balance	
	(a)	(b)	(c)	(d)	a-(b+c)	a-(b+d)
1969	481	394	263	142	-176	-55
1970	722	357	490	296	-125	69
1971	517	454	214	119	-151	-56
1972	404	386	167	64	-149	-46
1973	593	310	307	143	-24	140

The last two columns in Table 4 show the yearly difference between precipitation (input) and drain-flow plus evapotranspiration and, precipitation and surface flow plus evapotranspiration. We will not attempt to give a full explanation for the large difference in water balance between the drained and the undrained area. Table 6 is also relevant and helps with an explanation.

Table 6. Numbers of daily events of precipitation, surface run-off, and drain out-flow for the years 1968 to 1973. A year runs from Nov. 1 to Oct 31 inclusive.

Year	Precipitation events	Drainflow events	Surface-flow events
1968-1969	262	365	222
1969-1970	300	365	276
1970-1971	263	365	187
1971-1972	263	365	172
1972-1973	272	365	228

Drain-flow occurred on every day during the five years of measurements, while surface run-off occurred on fewer days than rainfall. The installation of a drainage system about 0.7m deep with consequent lowering of the watertable created a large sink surrounded by an extensive area of peat with a higher water table. This resulted in inflow to the drained area, and caused increased and continual drain-flow . Another but smaller source of extra water would have been dewatering of the drained peat which manifested itself as shrinkage but was not measured. In the last column of Table 4 surplus precipitation not removed as surface flow or evaporation is given. This appears to be seepage, which finds its way through the peat mass to other outlets or to greater depths. A general cross movement of water through the peat in this general area was reported by Galvin and Hanrahan (1967).

WATER RELATIONSHIPS IN A CONIFER FOREST

A similar investigation was also undertaken in co-operation with the Department of Forestry on similar peat on a site adjacent to Glenamoy. This study ran from 1971 to 1980 inclusive. At the start the trees were about 1.5m tall and measurements continued until the trees were at full canopy. The main purpose of the study was to determine the changes caused by the growing forest in peat/water relationships. Among the records that were kept were precipitation, drain flow, soil moisture tension, peat moisture content, peat shrinkage and watertable levels. Plots were approximately 1.25 ha in area. In the data presented here Plot A had shallow drains approximately 0.25m deep, and Plot B had drains approx. 0.75m deep. All drains were spaced at 2 m. The principal results are summarised here. Considerable difficulty was encountered in measuring flows, and occasional leakages were found in plot perimeters. Because of this the data are not presented as giving a precise water balance but it is considered that they provide a good picture of the overall effects. In Table 7 yearly data are presented on precipitation, outflow and water balance.

Table 7 Summary of Yearly Water balance (1971-1980) inclusive.

Year	Ppt	Outflow	mm	(Ppt - Outflow) mm	
	mm	A	B	A	B
1971	1170	605	638	565	532
1972	1232	631	567	601	665
1973	1382	707	673	675	709
1974	1508	751	684	757	825
1975	1172	465	426	707	746
1976	Not Available				
1977	1527	658	645	869	882
1978	1577	655	651	922	926
1979	1541	631	681	910	860
1980	1520	640	671	880	849

Data for 1976 is not available. A long dry spell in late summer caused cracking in the peat in the vicinity of the recorders and measurements were discontinued until repairs were made after the peat re-wetted.

The two main points evident in Table 7 are:

(a) Flows from both areas are much closer in value than the corresponding Glenamoy data. This is attributed to the larger plot size and the location of the plots in the landscape - at the top of a very gentle slope. Both factors reduced the danger of the sink effect ascribed to Glenamoy.

(b) When account is taken of the greater rainfall in the later years of the observations it is apparent that there is a decrease in outflow relative to precipitation, but a tendency towards stablization as full canopy is achieved. Less of the precipitation emerges in drainflow as the trees get larger. In the absence of other explanations it appears that more water is being evaporated - a phenomenon also reported by Law (1956). The trend is also shown in graphical form in Figure 7.

Watertable: The behaviour of the average watertable for both areas is shown in Figure 8. As expected, watertable falls in summer and rises in winter. The winter watertable in the deeply drained area is approximately the same as the summer watertable 0.4m in the shallow drained area. Summer watertable in the deeply drained area is about 0.7m i.e. the depth of the drain invert.

Subsidence: A graph of surface subsidence is shown in Figure 9. Subsidence is regular and continuous. It is more pronounced in the drained area, totalling about 0.4m over 10 years, as against about 0.2m in the shallow drained area.

Environmental Studies: As an adjunct to the hydrological study continuous sampling of drain water was carried out, in both grassland and forest. There was no evidence that any applied fertilizer was removed in drainage water at normal farming or forestry application rates. However when rates were increased to about three or four times normal considerable amounts of Potassium and Phosphorus emerged in the drainage water.

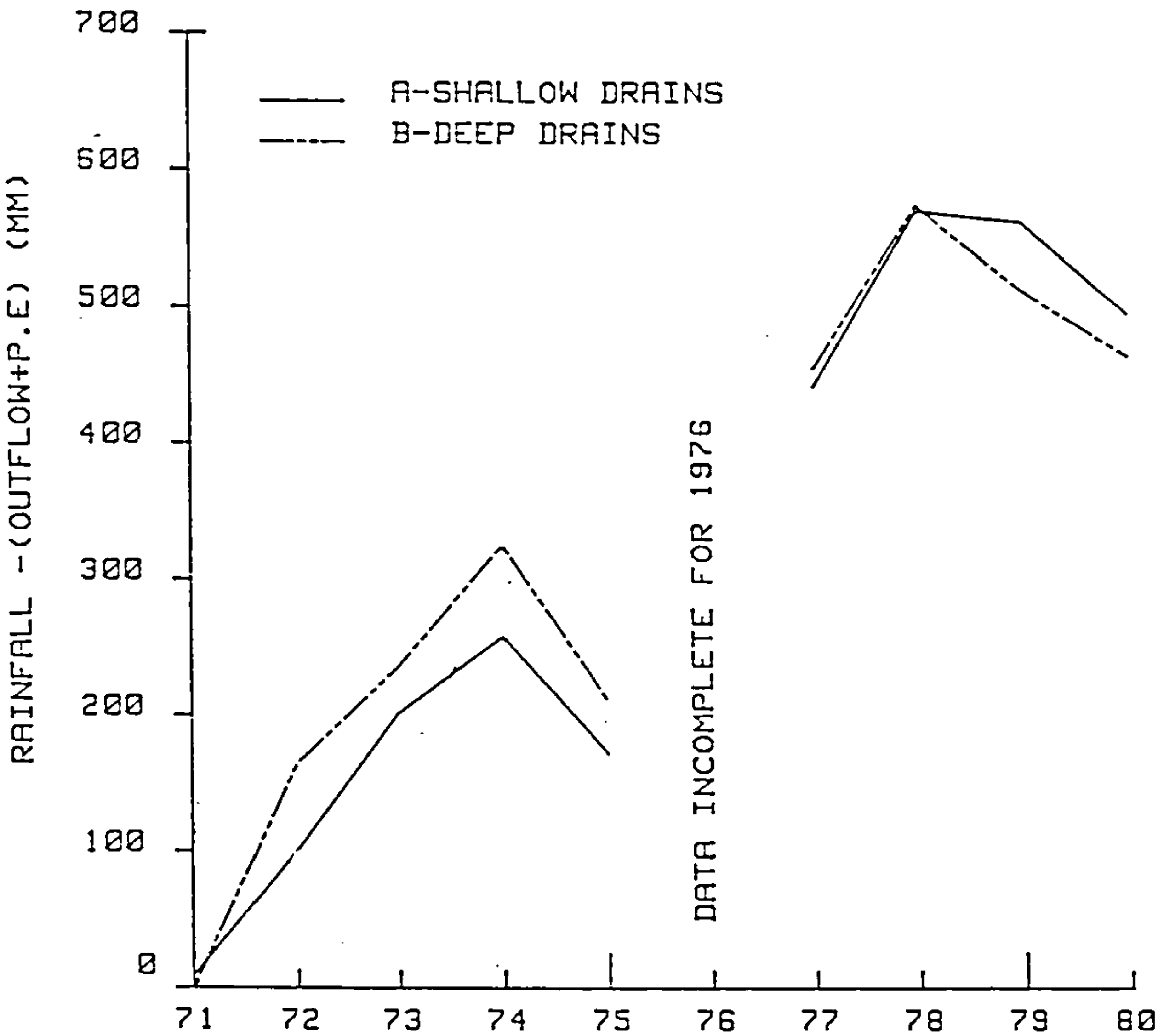

Figure 7 Annual changes in rainfall (outflow + P.E.)

Conclusion: The data indicate that:

(a) Evaporation from the forest increased as the forest developed and appeared to stablize at full canopy.

(b) Drainage caused subsidence in the peat and the subsidence was greater in the deeply drained area.

(c) Winter watertable in the drained area was on average as low as summer watertable in the undrained area.

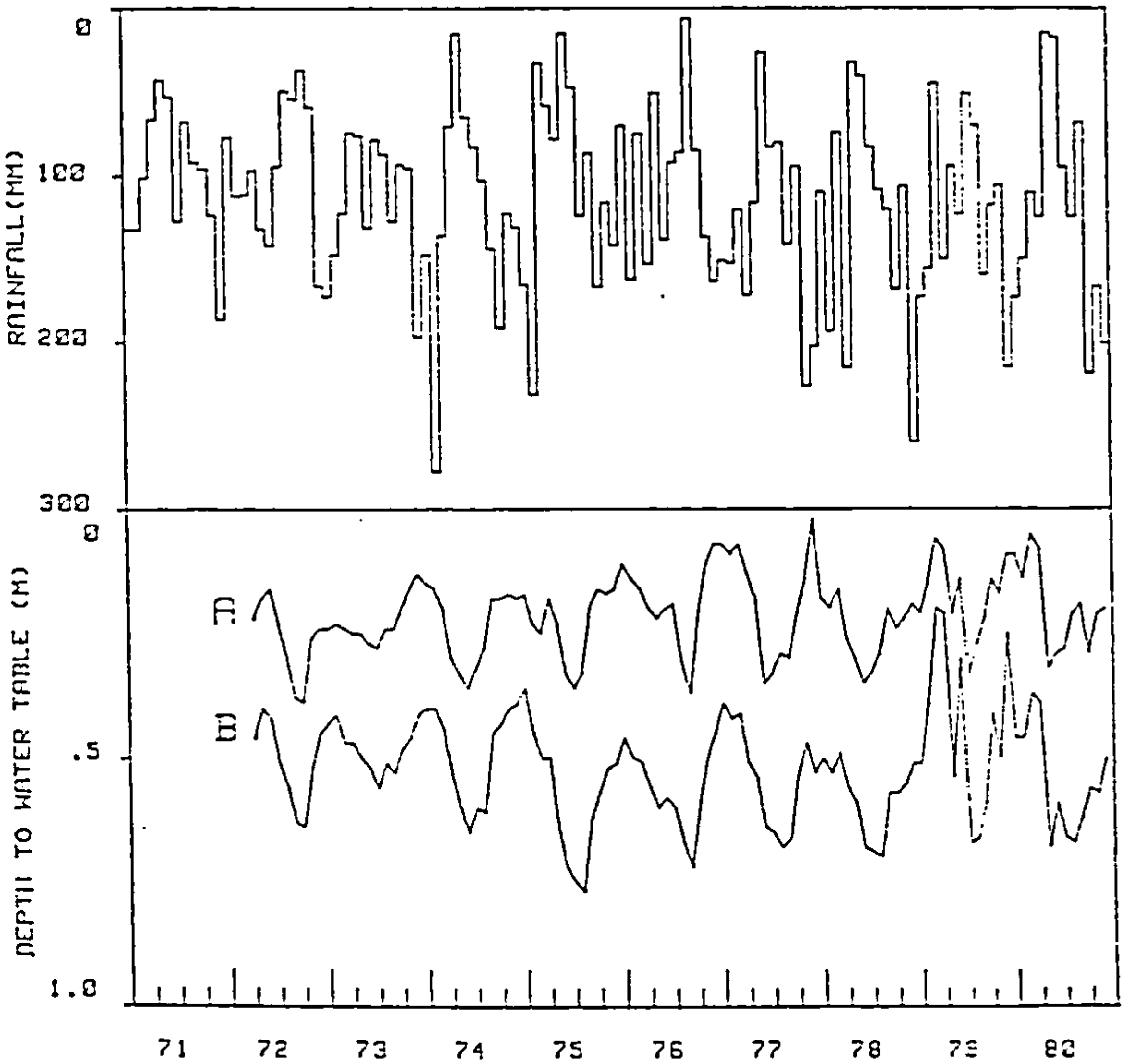

Figure 8 Average Watertable levels in the shallow drained (A) and undrained peat (B)

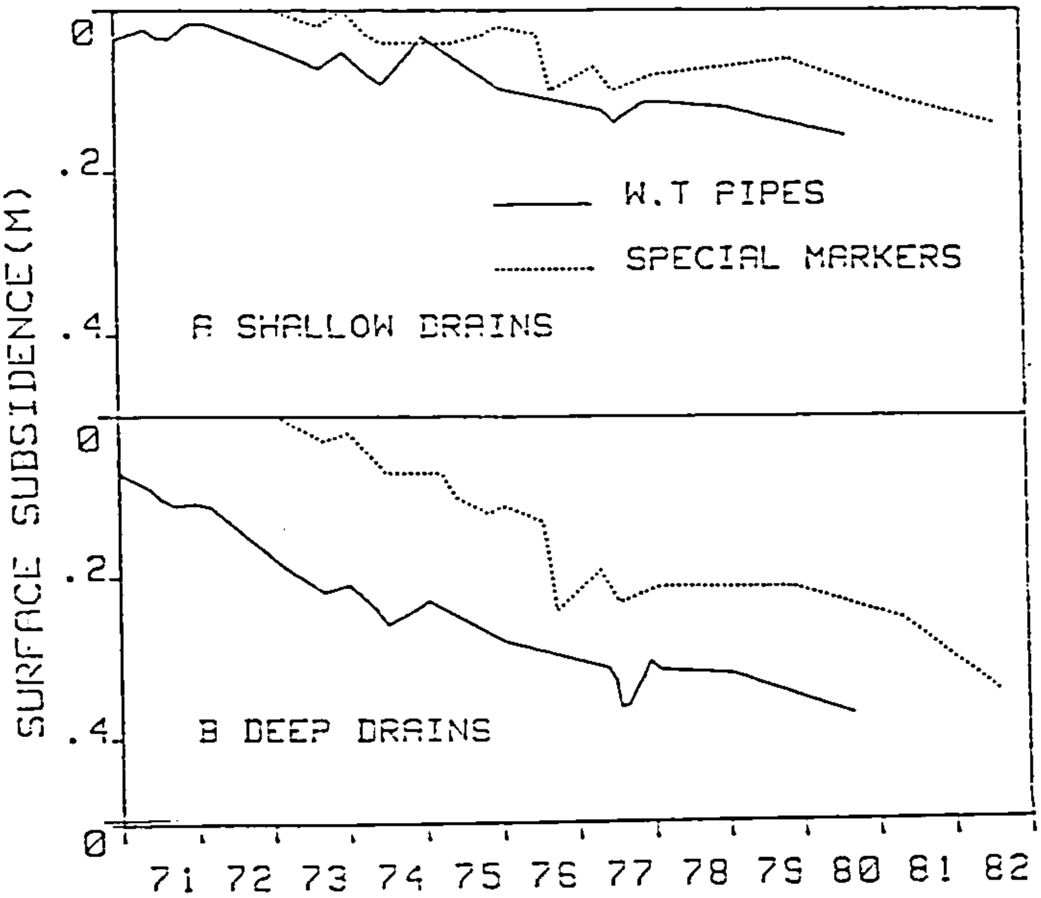

Figure 9 Surface subsidence from 1971 to 1982

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THE HYDRODYNAMICS OF RAISED BOGS : AN ISSUE FOR CONSERVATION

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Abstract

A raised bog is a complex ecological system which depends for its survival as a dynamic entity on the ambient hydrological and hydrogeological conditions. The conservation of such a system depends on an understanding of that hydro(geo)logy and its relationship with the ecology. The joint Irish-Dutch studies of the bogs at Clara and Raheenmore have concentrated on the hydrology and ecology of the peatland systems with a view to describing a strategy for their management. The raised bog at Clara has rare ecological features in its soak systems, which appear to be sustained by sources of water arising from convergent surface (acrotelm) drainage in a depression or mineralised water from convection calls in upwelling regional groundwater. Nevertheless, the raised bog itself appears to be almost hydraulically independent of the underlying regional groundwater system, separated from it by a layer of lake clays.

Hydrogeological characterisation of the bog system showed that the peat stratigraphy was reflected in the measured values of hydraulic conductivity and storage. The fen and bog peats forming the catotelm were of significantly lower hydraulic conductively than the growing acrotelm layer which encouraged rapid runoff at high water levels. Measurements of hydraulic conductivity in the peat confirmed the non-Darcian dependence of that parameter on hydraulic gradient. Numerical modelling of the bog system showed that, at the larger scale, the values of hydraulic conductivity were significantly higher than the measured values, reflecting the presence of macropores (cracks and subsurface drains). In the steady state, maintenance of water level, rather then flows, is the key requirement of conservation.

In the long term, however, drainage and peat cutting has and will cause subsidence of the bog surface with a lateral influence of several hundred metres from the lines of drainage. Reconstruction of the bog surface at Clara was based on the comparison of measured properties of peat to those of an undisturbed peat column. Results showed that, along a bog road, drainage has caused the peat to compact by nearly 50% in 200 years, although the rate of consolidation is decreasing logarithmically with time. Developed relationships between hydraulic conductivity, density and hydraulic gradient will provide the basis for future needs in modelling the long term effects of human activities on bog conservation.

Peat and Water

"The ground itself is kind black butter

Melting and opening underfoot"

(From: Boglands by Seamus Heaney, 1989)

To anyone who has worked barefoot when cutting peat, Seamus Heaney superbly evokes the physical quality of peat, a sensual quality that is dictated largely by its water content. Peat equals water....almost! It contains more water than milk - it is over 90% water by volume. Five metres of fibrous peat may contain 4.7m of water and as little as 0.3m of solid plant matter (Hobbs, 1986). The high water content makes it a very unusual, interesting but difficult geological material. Water in peat, as in other rocks, is present as free water in the pores, but unlike other rocks water is an essential part of the structure of the peat, bound physically, chemically, colloidally and osmotically. Only a small proportion is mobile, although this varies with the hydraulic conditions and with the physical state of the peat. Consequently,conventional hydrogeological parameters such as Darcian permeability and the coefficient of storage may not be appropriate.

Peatlands are accumulations of waterlogged plant organic matter topped by a surface layer of living plants. Water plays a vital role in peat formation, since it acts as a preservative against decay by excluding the entry of oxygen. A high water table is an essential condition for peat formation. Achieving this depends on a number of conditions, such as climate, topography, hydrology, geology and hydrogeology; conditions which vary throughout Europe. As a result, peatland distribution and types vary.

In Ireland there are three types of peatland - fens, raised bogs and blanket bogs. The main water source for fens is mineral-rich (usually calcium-rich) groundwater and surface water and so they have a neutral or slightly alkaline pH. Rainfall is the only source of water for bogs and as a result such peatlands are acidic and poor in minerals (ombotrophic). As such, "bog" is the accepted term to describe that category of mire or wetland whose only inflow is rainfall (Dooge, 1972). Raised bogs attained their finest development in Ireland, where they covered large areas of the Midland limestone plain. Most raised bogs in Ireland have developed from fens - as the peat continued to accumulate, it formed a flattened dome, slightly higher than the surrounding area and hence the name. The peat of raised bogs consists largely of water and sphagnum mosses, and can reach thicknesses of 15m. Blanket bogs form where rainfall exceeds 1200mm and falls on more then 250 days per year.

Raised bogs have two functionally different zones from a hydrological viewpoint - the acrotelm and the catotelm (Streefkerk and Casparie, 1989). This zonation is critical to the understanding of the hydrodynamics of raised bogs. The acrotelm is the system of living sphagna and other bog plants, and includes the water table which provides the water supply. It has a thickness of 0.1 - 0.7m, a high permeability, a large capacity to store water and it can shrink and swell. For a raised bog to function hydrologically, it must have an intact acrotelm over the majority of its surface (Kelly, 1993). The catotelm is defined as the hydrological system between the acrotelm and the mineral subsoil (Streefkerk and Casparie, 1989). It has a low permeability and a low specific yield.

The Role of Hydrology and Hydrogeology in the Conservation of Peatland

Peatlands were a major landscape feature and geological unit in Ireland as they covered 17% of the land surface - a higher proportion than in any other European country with the exception of Finland. Peatlands are worthy of conservation for a variety of reasons: geological, geomorphological, ecological, social and economic. However, the high water content, which has helped create an unusual ecosystem and landscape, means that peat is a delicate geological material prone to removal of water by drainage and consequently to damage and often destruction. Drainage, for instance, can stop peat formation, cause consolidation, shrinkage and contraction, and allow access of oxygen and wastage by biochemical oxidation.

Drainage, peat cutting, afforestation and agricultural development have destroyed most of the peatlands of north-western Europe, and have had a significant impact in Ireland. As a consequence, peatland conservation is now one of the most urgent nature conservation issues in Ireland and has been given a high priority by the National Parks and Wildlife Service (NPWS) of the Office of Public Works (OPW).

Bogs represent a delicate ecological equilibrium and are sensitive to the quantity and quality of water available. However, the ability of bogs to withstand changes in the water balance or water chemistry are factors which remain poorly understood. Also, the key element in the actual conservation of bogs is the understanding and control of the water regime - the water in, on, around and beneath them. Yet, until recently, scientists and engineers with expertise in hydrology and hydrogeology have had little involvement with peatland conservation.

Clara Bog: A Bog of International Importance

Clara Bog is a Nature Reserve and is a natural heritage area of international importance. It is located in County Offaly and is one of the largest (665 ha) remaining relatively intact raised bogs in Ireland. It has a well developed hummock-hollow system and is the only remaining Irish raised bog with well developed "soak" systems. The soak systems consists of areas of minerotrophic fen-like vegetation surrounded by ombotrophic Sphagnum mosses. These soaks are a particularly interesting ecological feature. At present the bog is in the form of two domes separated by a road with associated drains. Drainage and/or peat cutting has occurred on all sides of the bog and is still been undertaken particularly on the south side.

Irish-Dutch Raised Bog Geohydrology and Ecology Study

BACKGROUND

At present there are no intact raised bogs in the Netherlands, so attempts are being made to restore some of the peatlands. Ireland in contrast still has several raised bogs, although many are damaged or under threat from drainage and peat cutting. One of the policies of the NPWS is to conserve a number of Irish raised bogs. Following a meeting in 1987 between Mr. N. Treacy, Minister for State at the Department of Finance and Mr. G.J.M. Braks of the Netherlands government, it was decided to set up a joint Irish/Dutch research project on the geohydrology and ecohydrology of Irish raised bogs. Two bogs in County Offaly were chosen - Clara and

Raheenmore. The study was funded on the Irish side by the NPWS, with substantial assistance from the Geological Survey of Ireland (GSI).

OBJECTIVES OF STUDY

The overall objectives were:

- to develop sustainable management measures for the conservation of Clara and Raheenmore bogs and, by extension, other raised bogs in Ireland.
- to provide information to assist in raised bog restoration in the Netherlands.

Specific aims were:

- to investigate the various elements of the water balance such as rainfall, evapotranspiration, surface run-off, groundwater flow.
- to understand the hydrodynamics and hydrochemistry of raised bogs, and in particular of the soak systems.
- to understand the relationship between the bog plants and the water regime and chemistry.
- to determine the impact of drainage and peat cutting.

STUDY APPROACH

The conservation of a raised bog is dependent on an understanding of the water balance or hydrodynamics at a scale appropriate to the needs of the required management. Effects of peat cutting and drainage and the degree to which they should be allowed in a conservation area can only be ascertained on the basis of the same understanding of the water balance.

For an ombotrophic raised bog, as a hydrogeologically contained system, the water balance is described in the terms of effective precipitation (rainfall P less evapotranspiration E) and surface runoff R:

$$P + E + R = \Delta S$$

The change in storage ΔS in the bog system may be assumed as negligible over an annual cycle, but will assume significance in terms of long term drainage or in the growth of the bog itself. The low permeabilities and dynamic nature of the storage capacity give the water balance of a raised bog a scale of dependency in time as well as space.

The approach taken in the Irish-Dutch study has been to attempt to understand the "steady state" dynamics as they exist now, supported by an investigation of the evidence for longer term changes in the bog morphology caused by drainage and cutting. Simulation of the dynamic changes in morphology is a longer term objective.

The current state of the bog therefore can be characterised by measurement and estimation of the fluxes in the water balance, by assessment of the existing pattern, and ultimately by the use of a numerical simulation model to verify the conceptual hydro(geol)ogical model that emerged. Combined with the effects of drainage over the last 200 years, a strategy for progressive conservation is being devised.

The objectives of the study could only be achieved by detailed geological, hydrological, hydrogeological, hydrochemical and ecohydrological investigation; in other words by a multidisciplinary approach. This work included techniques such as geological mapping, geophysics, surface runoff measurements, rainfall measurements, evapotranspiration estimation, drilling, piezometer installation, water level monitoring, permeability determinations, hydrochemical sampling and analyses, numerical modelling and vegetation mapping. The geologists provided the local and regional framework for the bogs, the hydrologists and hydrogeologists provided the local and regional water flow regime and, together with the ecologists, integrated the water regime with the different plant communities. With the understanding of the water regime and the ecology provided by the integrated multidisciplinary approach, technical measures, such as water control by dams, have been initiated to enable either conservation or restoration of the peatlands. This approach also allows the conservation authorities to assess the conservation viability of a particular peatland prior to purchase or prior to ranking it as a Natural Heritage Area.

As a means of achieving an input from different disciplines, the following organisations contributed to the study:
OPW; GSI; Teagasc; Sligo RTC; TCD; UCG; Department of Nature Conservation, Environmental Protection and Wildlife Management (The Hague); National Forest Service or Staatsbosbeheer (Driebergen); Wageningen Agricultural University; University of Amsterdam and Imperial College London. Students from the third level colleges carried out much of the data collection and analysis, and the student reports provide the basis for the final assessment and decision-making. The study was managed by a Technical and a General Supervisory Group consisting of OPW, GSI and Teagasc staff and Dutch counterparts.

The overall assessment of the different elements of the study has not yet been completed. Consequently, the following comments and conclusions are provisional. Also, they are focused largely on Clara Bog.

Regional Geological and Hydrogeological Setting of Clara Bog

Clara Bog has formed in an elongated basin, which is surrounded on one side (northern) by a prominent esker sand and gravel ridge and on the other by glacial tills(Figure 1). The geological succession at the bog is as follows:

Lithology	Thickness (m)
Poorly humified peat	0 - 6.5
Strongly humified peat	0 - 3.5
Fen peat	1 - 3.5
Shell marl	0 - 0.4
Lake clay	0 - 7
Till	0 - 18
Limestone bedrock	-

(based on data from Smyth (1994) and Bloetjis and van der Meer (1992))

Most of the basin is floored by lake clay, although there are small ridges and hillocks of till directly beneath the peat in places. The greatest peat depth is 12m (Smyth 1994).

The hydraulic conductivities calculated for the mineral subsoils and the bedrock are as follows:

Lithology	Hydraulic conductivity (m/d)
Lake clay	$1x10^{-5}$ - $1x10^{-4}$
Till	$1x10^{-1}$ - $8x10^{2}$
Esker sand and gravel	$2x10^{1}$ - $5x10^{3}$
Limestone	$2x10^{-2}$ - $4x10^{1}$

(Flynn, 1993)

There are two distinct hydrogeological units at Clara Bog, separated by the lake clays(van der Boogaard, 1993; Flynn, 1993). Regional groundwater flow in the till and limestone beneath the clay is generally from north to south beneath the bog (Figure 2), with recharge occurring in the area of esker sands/gravels and discharge to perimeter drains and the Silver River south of the bog (Figure 1). In contrast, as the water table closely mirrors the topography, the groundwater flow direction in the peat radiates out from the apex of each of the two bog domes.

One of the issues in characterising the behaviour of a raised bog is the role of regional groundwater flow. In this case, a combination of evidence from the measurements of hydraulic conductivity, the potentiometric surface map and modelling, indicates that the bog is operating as an independent entity. However, at the fringes of the bog (the lagg zone), there is evidence in places of mixing of upwelling groundwater flow and surface runoff from the bog. (The vegetation in lagg zones is considered to be important from an ecological point of view).

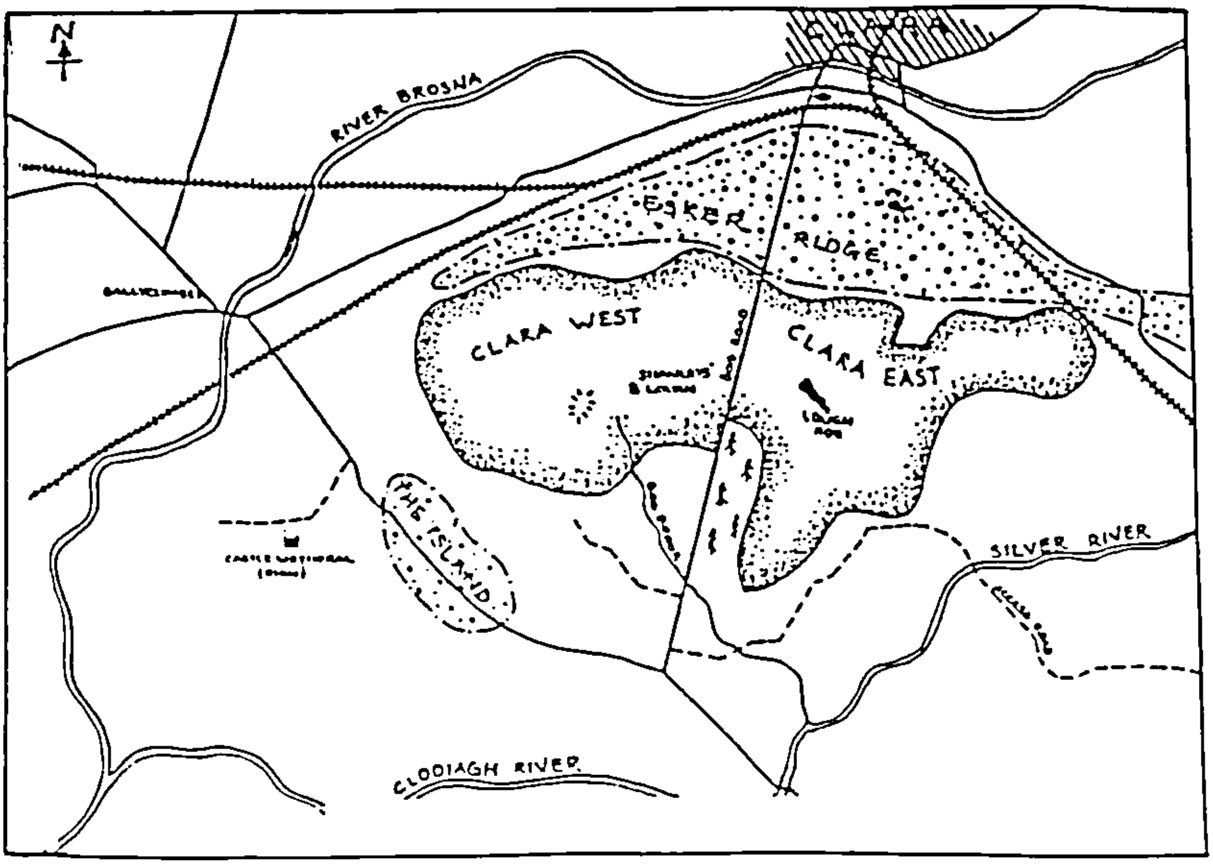

Figure 1 Location of Clara Bog, Co. Offaly (from Preston, 1993).

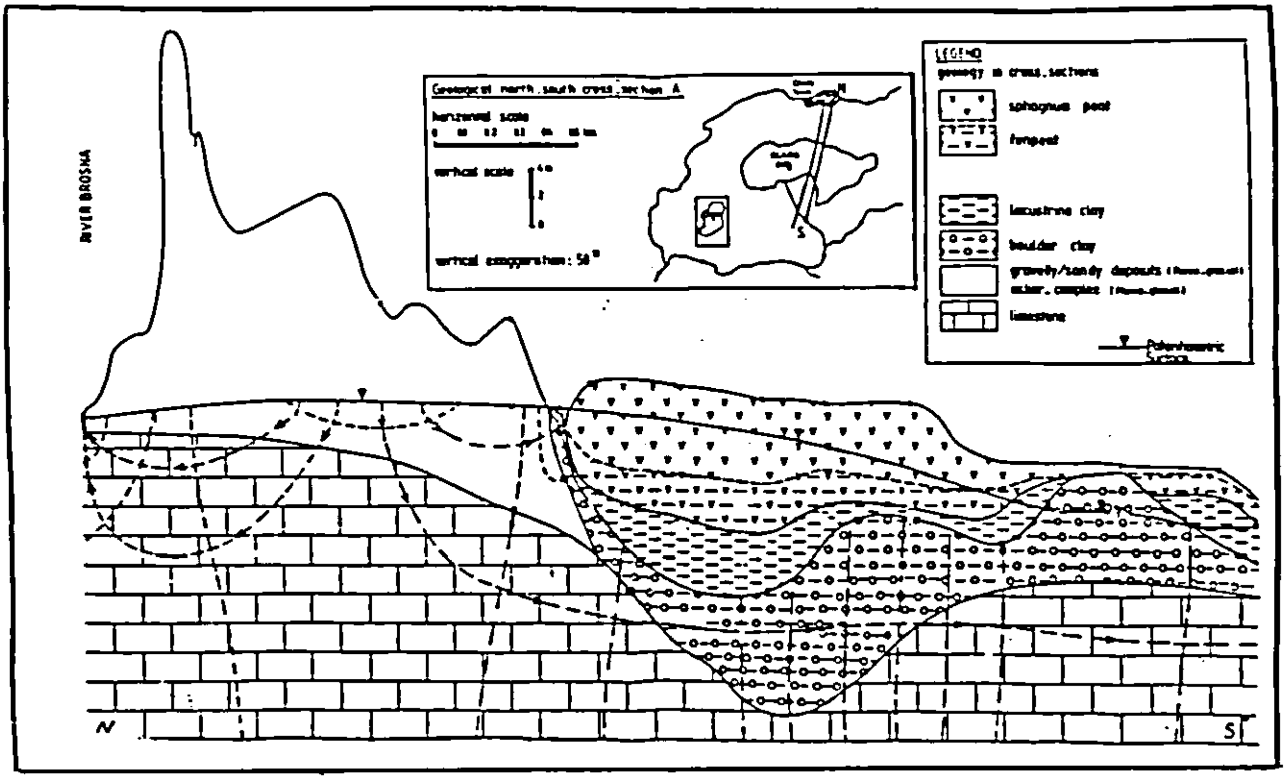

Figure 2 Hydrogeological cross-section through Clara Bog area (after Flynn, 1993).

Hydrogeological Characterisation of a Raised Bog

HYDRAULIC CONDUCTIVITY

Determining the hydraulic conductivity of peat is problematical as it can vary with the imposed head. A large imposed head gives higher values and visa versa. The reason for this has been attributed to non-Darcian flow (Ingram et al., 1974) and to compression or expansion of the peat matrix, which depends on the effective stress caused by addition or removal of water (Hemmond and Goldman, 1985). It was concluded that the constant head technique with a low imposed head is, theoretically, the best method of estimating hydraulic conductivity as it eliminates storage influences and has a low disruptive effect on the peat (Flynn, 1993) (ten Dam and Spieksma, 1993). However, the method proved exacting and difficult to implement. Consequently, it was decided that a rising or a falling head test could be used to give an idea of representative hydraulic conductivity, provided that data analysis was confined to periods when the imposed head was lowest. A statistical comparison of the results from the rising and falling head tests (Sytsme and Veldhuizen, 1992) showed that the falling head test produced hydraulic conductivity values with the least variance thus indicating a preference for the falling head test.

The hydraulic conductivity of the catotelm was found to be variable. Ten Dam and Spieksma (1994) found that at Raheenmore it ranged from 1×10^{-5}-4×10^{-1} m/d with a mean of 1.5×10^{-3} m/d. At Clara, 107 tests were performed giving values ranging from 10^{-5}-10^{0} (Leene and Tiebosch, 1993). Some of the lower values were obtained in areas affected by subsidence.

Following the recommendation of Sake van der Schafe of Wageningen University, the acrotelm was mapped and its transmissity was calculated. Hydraulic conductivities were found to be generally high, although they varied with the humification degree. The range was 0.1 - > 1000 m/d, with most values above 25 m/d.

This variability illustrates the scale dependence of hydraulic conductivity measurements in peat. These determinations used narrow (25mm) diameter piezometer tubes and thus were vulnerable to small changes in peat texture and density. The role of macropores (pipes and cracks), which are known features of raised bogs, could only be determined by larger scale evaluation techniques i.e. numerical modelling.

WATER BALANCE

An investigation of the hydrodynamics of a raised bog system is aided by the construction of an annual water balance, which enables identification of the key hydrological fluxes and the establishment of a conceptual model of the system. In an ombotrophic bog, the inflow is from precipitation. Outflows are evapotranspiration, surface runoff and losses laterally through the catotelm to peripheral drainage or regional groundwater beneath the bog. An important unknown may be the degree of interaction between the peat bog and the underlying regional groundwater system. Not only may there be a physical flow leakage in either direction, but the resulting mixing of higher pH limestone waters with the relatively acidic bog environment can have significant ecological consequences in terms of different vegetative systems (e.g. soaks, lagg zones).

At Clara and Raheenmore bogs, rainfall was measured on-site over a period of two years by University of Wageningen students using recording tipping bucket and manually read daily gauges. Although the tipping bucket gauges consistently under-registered compared to manual gauges, probably due to instrument flooding in higher intensity rainfalls, rainfall was relatively accurately observed on the bogs.

Evapotranspiration from a wetland is a key component of the water balance and the most difficult to determine in practice. On a large exposed Midland bog, such as Clara, wind fetches are long, and evaporation may occur at near open water rates when levels are close to surface and evapotranspiration occurs from the vegetation itself. Therefore, although hydrometeorlogical data was available from Meteorological Service stations at Birr and Mullingar (each over 30km away), it was decided to measure evapotranspiration on-site using a variety of approaches. An automatic weather station was established at Clara Bog measuring and recording wind speed and direction, solar radiation, net radiation, air temperature and relative humidity, at hourly intervals. To complement these data, fifteen weighable buckets (400mm diameter, 500mm depth) lysimeters were installed adjacent to the weather station (Leene and Tiebosch, 1993). Five vegetation types were established and direct measurements of evapotranspiration was made weekly over a 10 month period in 1992-93. Estimates of monthly evapotranspiration have also been based on the methods of Penman, Penman-Monteith, Thom and Oliver and Morton (Leene and Tiebosch, 1993; Maina, 1993; Shuaib, 1993). Lysimeter data indicated an annual evapotranspiration loss of 587 mm at Clara with little difference between species of vegetation over the year. All the other methods yielded considerably lower annual values (see Figure 3), mainly underestimating the measured losses in the winter period. In general, the complementary relationship of Morton (Maina, 1993) gave the closest unconstrained predictions of the reference lysimeter measurement. However, Shuaib (1993) using the Penman-Monteith equation, achieved good predictions of the lysimeter values by seasonally optimising the canopy resistance factor. The exigencies of operating lysimeters in difficult bog conditions, particularly in winter, on the other hand may have produced overestimates of evapotransation but nevertheless, it would appear that evaporative losses are significantly more than might have been estimated using data from a regional, conventionally sited Meteorological Service stations.

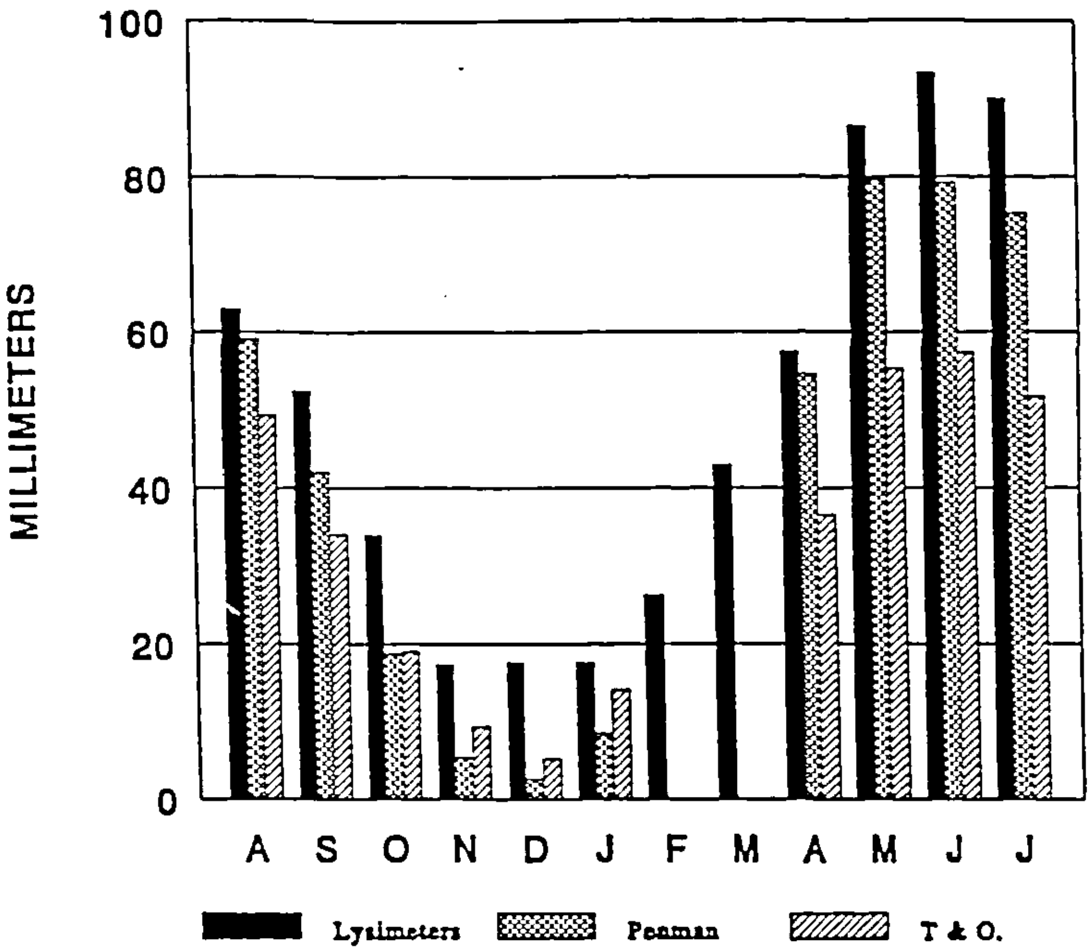

Figure 3 Monthly estimates of evapotranspiration on Clara Bog, 1992-1993, from lysimeters, and the methods of Penman and Thom and Oliver (from Leene and Tiebosch, 1993).

Surface drainage from a bog system is usually significant in the wetter months when the water level in the peat is near the surface and runoff is relatively rapid. Defining the catchment area on the peat from which runoff is collected in an effluent (or gaining) drain is problematical with the hummocky bog surface and small topographic gradients. At Clara and at Raheenmore, weirs and recording flow gauges were installed on the principal streams from which long term estimates of bog discharge were made. Perimeter drains collect runoff from both the bog and the surrounding mineral soils and partition of component flows was achieved using hydrochemistry.

Leakage through the clays underlying the peat may be estimated using observations of water levels in sets of vertically distributed piezometers in response to rainfall, but such measurements are inevitably localised. Using this approach, some 2% of incoming rainfall at Clara was estimated to be lost through leakage. Although the hydraulic gradient is clearly downwards from the peat into the regional groundwater system, there is considerable uncertainty in the magnitude of the flow given the apparent low hydraulic conductivities of the clays and the catotelm peat. Not withstanding the considerable uncertainties in the determinations of the hydrological fluxes, the annual water balance for the raised bog at Clara has been estimated by Leene and Tiebosch (1993) as:

Precipitation	922mm
Evapotranspiration	587mm
Vertical seepage	14mm
Drainage discharge	300mm
Change in storage	20mm

In this case the surface catchment area was treated as a residual because of the uncertainty in delimiting it. Change in storage in a peat bog must take account not only of a measured change in water level in the acrotelm (estimated specific yield of 20-30%) but also of the physical change in the structure of the peat on an annual basis. Benchmark measurements at Clara indicated that the bog surface could swell or rise by as much as 70mm during the wetter period

Precipitation may increase during the winter slightly but evapotranspiration is much more strongly seasonal (Figure 4) and dependent on water level. Drainage discharge is also seasonal (but more controlled by rainfall events), although proportional runoff is controlled by water level - for a given rainfall in high water table conditions there is less available storage in the acrotelm and therefore there is a greater runoff. Vertical seepage is negligible and probably at a steady rate. Initial analysis of the water balance indicates, as expected, that maintenance of the water level in the bog, rather than a particular flux rate, is the critical control.

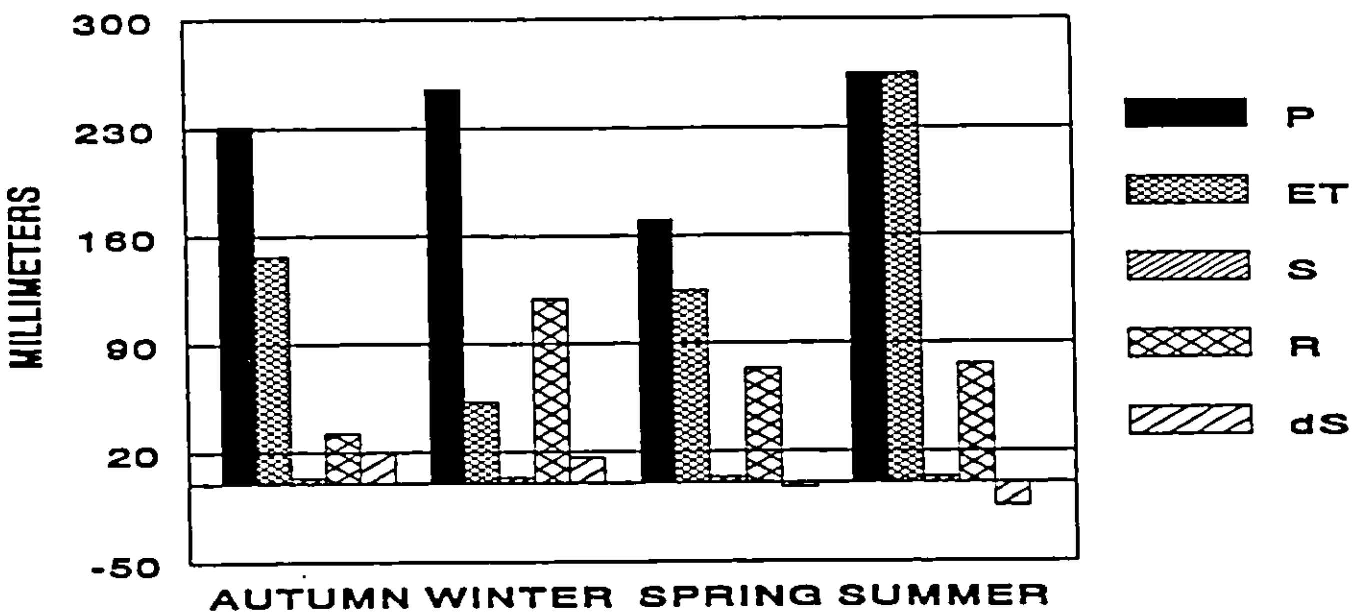

Figure 4 Estimated seasonal water balance for Clara Bog from August 1992 to July 1993. P = precipitation, ET = evapotranspiration, S = vertical leakage, R = surface runoff, dS = change in storage (from Leene and Tiebosch, 1993).

An analysis of this water balance and the consequent hydrodynamics has been undertaken using a numerical model of sections of a bog and of the bog itself. The three-dimensional finite difference model, MODFLOW, developed by the US Geological Survey, was applied to the measured data to determine the validity of the conceptual model of a raised bog existing as a more or less independent system (Blackwell, 1992; Preston, 1993). The model was used in inverse mode to determine the hydrogeological parameters necessary to reproduce the observed water levels as measured by piezometry. Account was taken of the known peat stratigraphy (fen peat and bog peat in the catotelm and overlying acrotelm). At Clara, modelling was undertaken on the basis of the known piezometric surface in the peat and in the underlying limestones and of the known drainage discharges. The results indicated that the losses through the base of the bog was likely to be small but the hydraulic conductivities in the catotelm peat was likely to be much higher when evaluated at the bog scale than at the scale of individual piezometer measurements. Mean hydraulic conductivities for the catotelm peat of the order of 0.1m/d and for the upper catotelm layer of 0.5 - 3m/d are required to reproduce the observed water levels and flows in the bog. Although these values are significantly higher than the individual measurements, they indicate strong effects of macropores (cracks and pipes) at the larger scales.

ORIGIN OF SOAK SYSTEMS

Soak systems add significantly to the diversity of plant species on a raised bog and are now extremely rare features (Kelly, 1993). Consequently, their conservation is an important objective. However, successful conservation depends to a significant degree on understanding their origin. Three main soak systems and three minor areas have been recorded on Clara bog (Kelly, 1993). Two have been examined in some detail - Shanley's Lough and Lough Roe (Figure 1).

The origin of the mineralised water that supports the minerotrophic vegetation at Shanley's Lough is related to the bog morphology. Shallow depressions resulting from subsidence have focused relatively large volumes of normal raised bog water (low electrical conductivity (EC), NaCl water type and Na:Ca ratios >>1), (Flynn, 1993; Kelly 1993). The origin of mineralised water in Lough Roe is still not known with certainty. Unlike Shanley's Lough, Lough Roe has formed near the crest of a dome, probably as a result of a tension crack. The water chemistry is different with higher EC's, CaCl and $CaHCO_3$ water types, and Na:Ca ratios <1. A suggestion by Daly (1987) that upwelling regional groundwater flow was the source of the minerals was not supported by hydraulic head data in the area, which showed that flow was downwards and not upwards. Streefkerk (pers. comm.) suggested that convection currents could be bringing the mineral-rich water to the surface. Flynn (1993), who assessed earlier work by Schaffers and van der Meer (1993), concluded that convective currents do occur, but only in the colder months. Thus, regional groundwater could be drawn to the surface through a "window" in the lake clays by convective currents and may be providing the minerals necessary for the Lough Roe vegetation.

Drainage and Subsidence

A water balance combined with numerical models of the fluxes in the system necessarily represent the hydrodynamics of a raised bog as they currently exist. Peat cutting and drainage along bog roads or around the bog perimeter will have longer term effects involving subsidence of the peat and consequent changes in the flow regime and ecology. Issues for conservation are not only whether the status quo can be maintained but also whether the longer term effects can be arrested or even reversed.

Subsidence in peat occurs as a result of removal of water from its structural matrix. This reduction in thickness may occur as the result of loading the surface or by directly reducing the subsurface pore water pressure by drainage. At Clara, a road runs between the two domes of the bog and is suspected of being the cause of this morphology. Moreover, the soak of Shanley's Lough may have been created by long term drainage at peat cuttings causing subsidence of peat in a depression, thereby focusing more runoff water into one area.

The road across the bog at Clara was built some 200 years ago as a track, probably laid initially on a mat of branches and it still retains a thin gravelly sub-base. As such, the road itself combined with very little traffic represents a light loading of the peat. However, open drains either side of the road, up to 2m deep, represent direct surface and subsurface drainage from the bog to the nearby Silver River.

Several investigations (Bell, 1991; Samuels, 1992; ten Dam and Spieksma, 1993) have attempted to evaluate the degree of subsidence at Clara resulting from drainage associated with both the road and of the peat cuttings and to determine its lateral extent. The approach has been to compare the properties of the peat in relatively undisturbed, uncompacted columns with those properties measured nearer the lines of drainage. The consolidation ratio (length of a consolidated peat column/length of an unconsolidated peat column) can be shown to be equivalent to the ratio of the average volumetric concentration of organic matter in the same columns, modified by an estimate of any oxidation of organic matter that might have taken place. This relationship formed the basis of the ten Dam and Spieksma's (1993) reconstruction of the bog surface across the soak at Shanley's Lough. Bell (1991) and Samuels (1992), respectively, used the measurements of peat densities and of the ratios of the mass of water to solid material relative to similar measurements in an unconsolidated column, in order to reconstruct the bog surface across the central road at Clara (Figure 5). In addition, a 1910 level survey of the road was compared to current levels to measure actual subsidence of the road over the last 80 years. These reconstructions generally indicated that the original bog surface was at or above the elevations of the current dome crests thus showing a compaction of some 5m in an original peat column of 11m in the vicinity of the road. Effects of subsidence were evident for over 500m on either side, the compaction having occurred over the 200 years of the road's existence. However as settlement has a logarithmic relationship to time, the rate of settlement at the road has slowed to a predicted 0.5m in the next 200 years.

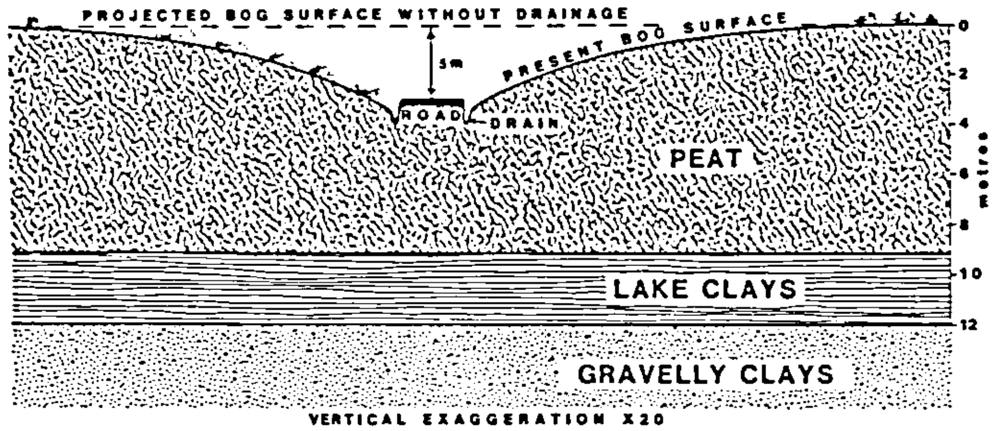

Figure 5 Diagrammatic cross-section across bog road at Clara.

The implications for the hydrogeology of the bog is that firstly, the acrotelm and its ecology cannot be sustained along the lines of the drainage and secondly, the hydraulic conductivity of the catetolm decreases significantly under compaction and drainage. A relationship for predicting hydraulic conductivity as a function of density and hydraulic gradient was developed (Samuels, 1992) for Clara, which is a necessary part of the future development of a model for predicting the effects of human activities on bog conservation.

Vegetation mapping (Kelly, 1993) has shown that the wet areas are now confined to the core and the soaks. Where the bog surface gradient is more than 0.3m/100m, due to subsidence, the acrotelm is badly damaged or absent (Streefkerk, pers. comm.). Consequently a significant proportion of the acrotelm has been degraded by drainage. As the survival of the bog depends on the acrotelm, the damage at Clara is significant.

Remedial Measures

The OPW, with assistance for Staatsbosbeheer, are now considering engineering measures firstly, to halt the degradation of the bog and secondly, to restore the natural systems in some of the damaged areas. Such measures may be achieved through the use of peat embankments and dams,and closure of drains to create hydraulic boundary controls in the form of lakes or ponds. Experimental dams are being constructed on Raheenmore bog as a means of restoring the acrotelm. Measures involving the regional groundwater situation may be needed if the lagg zones are to be restored and possibly also if some of the soaks are to be conserved in the long term.

Conclusions

1. Peat is an unusual geological material in that it consists of over 90% water by volume; water that is an essential part of the structure of the peat. Also, only a small proportion of this water is mobile, with the amount varying with the hydraulic conditions and with the physical state of the peat.
2. A key element in the conservation of bogs is the understanding and control of the water regime.
3. Assessment of the water regime of bogs and the relationship to vegetation requires a multidisciplinary approach.
4. Determining the hydraulic conductivity of peat is problematical as it can vary with the imposed head. It was concluded that although the constant head test is, theoretically, the best method of estimating hydraulic conductivity, the falling head test can be used to estimate the representative hydraulic conductivity, provided that data analysis is confined to periods when the imposed head is lowest.
5. From a hydrogeological viewpoint, raised bogs can be divided into two functionally different zones - the acrotelm, which forms a thin (0.1 - 0.7m), high permeability, saturated top layer and the catotelm, which is a thicker (up to 14m), low permeability layer between the acrotelm and the mineral subsoil.
6. Evapotranspiration losses from the bog surface were found to be significantly more than estimated using data from a regional, conventionally sited Meteorological Service station.
7. Defining catchment areas on raised bogs is difficult due to the small topographic gradients and the hummocky bog surface, thus causing difficulties in water balance assessments.
8. The ability of the bog to swell by as much as 70mm means that surface runoff is reduced while the storage created by the swelling is available and is then rapid subsequently in the high permeability acrotelm.
9. Using both modelling and leakage estimates from hydraulic conductivity and head data, the main body of Clara bog was shown to exist more or less independently of the regional groundwater system.
10. Soak systems are rare ecological features which appear to be sustained by sources of water arising from convergent surface (acrotelm) drainage in a depression or mineralised water from convection cells in upwelling groundwater.
11. Modelling suggested that the bulk hydraulic conductivity of the catotelm is greater than implied by individual measurements, due probably to the presence of macropores (cracks and pipes).
12. Subsidence of up to 5m, due to drainage, has had a major impact on Clara Bog. In particular, drainage associated with the road crossing the bog has converted the original bog dome into two domes and has caused subsidence for over 500m on either side. Where the bog surface gradient is more than 0.3m/100m, due to subsidence, the acrotelm is badly

damaged and consequently the bog is not functioning properly either hydrologically or ecologically in these areas.

13. A significant proportion of the acrotelm on Clara Bog has been degraded by drainage.
14. A better understanding of the hydrodynamics and prediction of the long term effects of peatcutting and drainage on the sustainability of the bog ecosystem may best be approached through further numerical modelling. Thus the effects of different conservation measures may be assessed.

Acknowledgements
This paper represents the authors' overview of a comprehensive multidisciplinary investigation of the hydrodynamics and ecohydrology of Clara bog and Raheenmore bog. The paper draws on work from a large team of scientists and engineers in the institutions involved from Ireland, the Netherlands and England. The Technical Supervisory Group for the study included J. Streefkerk (Staatsbosbeheer), J. Ryan (OPW), B. Hammond (Teagasc), J. Martin (OPW), Sake van der Schafe (Agricultural University of Wageningen), M. Schouten (Staatsbosbeheer) and W. P. Warren (Geological Survey of Ireland).

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APPLICATIONS OF THE WATER BALANCE

WATER BALANCE OF WETLAND AREAS

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Abstract

The water balance is central to the study of wetland hydrology, as the relative significance of the various inputs, outputs and storages underlies the hydrological functions, water quality functions and the management of wetlands for conservation. This paper presents a brief review of the water budget, followed by an overview of three important areas of interaction between the wetland and its surroundings: the relationship between the wetland and rivers (the regulation function), the wetland and the atmosphere (wetland evapotranspiration) and the wetland and groundwater in aquifers. Research in these areas is still needed for the wise management of wetlands, and a full appreciation of their function and value.

Introduction

In the world's wetland areas, ecology, microclimate and the local economy depend on excess water. Wetland soils are at least periodically saturated, and wetland plant communities are often dominated by hydrophytes, plants that are adapted to the often severe problems posed by saturation, particularly oxygen depletion and the toxic products of anaerobic decay. Many natural wetlands are highly productive in the ecological sense, with a net primary productivity almost the same as that of rice fields and rather more than that of non-wetland vegetation (Aselmann & Crutzen 1989), but the fruits of this productivity have been under-valued, even despised, by man, and a sustained campaign of drainage, "reclamation" and non-sustainable exploitation has left wetland areas severely depleted (Bulavko 1971; Dahl 1990; Dahl & Johnson 1991; Denny 1993). Belatedly, the value of natural wetlands has been appreciated, but the hydrological processes defining wetland habitats, and the possibly beneficial effects of wetlands on the hydrological cycle, i.e. what might be described as their "perceived functions", are as yet poorly understood. There is room for much more research in:

(i) the relationship of wetland hydrology with wildlife, and with the related challenges of conservation, restoration and creation of wetland habitats

(ii) the interactions of wetlands with groundwater and surface water, vitally important for assessing wetland contribution to control of water quantity and quality. Where the habitat argument is not considered strong enough, it may be that the hydrological functions of wetlands are sufficiently important to justify conservation and management rather than destruction (Larson 1982). Wetlands are increasingly being regarded as natural water treatment facilities, but they will require careful management if we are not to store up problems for future generations.

The key to understanding the workings of wetlands lies in the water budget, which expresses the movement of water into, out of and through the wetland, and the storage of water within it. The water balance is central to hydrology: it is not surprising, then, that the water balance of wetlands is a large subject, touched on rather than covered by this paper. Four main subject areas are addressed here: the use of the water balance in wetland studies, wetlands and rivers, evapotranspiration from wetlands, and the interactions between wetlands and groundwater. The last three aspects encapsulate many of the most challenging problems facing wetland hydrologists today.

The wetland water balance

The wetland water balance attempts to take account of all inputs, outputs and storages in the hydrological system, i.e. the terrestrial phase of the water cycle. Unlike a drainage basin, however, the wetland is not delineated by convenient no-flow boundaries, and its water budget must include terms to represent groundwater and surface water inflows as well as outflows (Dooge 1975; Gilman 1994).

For any time interval, the full water balance for a wetland site can be expressed in the form

$$P + G_{in} + Q_{in} = E + G_{out} + Q_{out} + \Delta s$$

where P is precipitation
G_{in} is the groundwater inflow
Q_{in} is surface inflow
E is actual evaporation from the wetland
G_{out} is the groundwater outflow
Q_{out} is the surface outflow
and Δs is the change in water storage, usually seen as a change in water level or the water table.

In Figure 1 the wetland system is further broken down to show the various components of flow and storage, not all of which will necessarily be present or highly significant at all sites.

Many factors, from geology to climate, contribute to the formation of a wetland (Winter 1988). Local topography defines the initial location of a wetland, even if with later development, as in the case of blanket mire, the wetland becomes virtually self-sustaining. The universal requirement for excess water imposes some uniformity on the overall appearance of the wetland landscape, typically with expanses of low-growing moss, sedge or dwarf shrubs, shallow water bodies with fringing vegetation, grassy plants growing as emergents and highly adapted hydrophytic trees, but the range of wetland types is such that attempts to summarise wetland hydrology, like lost expeditions, have often disappeared into a morass of terminology and classification systems. It is intended here to give a few examples of studies in which a full wetland water balance has been attempted, and to examine the way in which this information has been used.

The greatest concentration of wetland is in the Arctic and subarctic zones,: for instance the areas of wetland in the former USSR and Canada are 1.51 and 1.27 million sq.km respectively (Aselmann & Crutzen 1989). Precipitation in high latitudes is low, and wetland development is possible because of two related climatic factors, low temperatures which inhibit decay and promote peat growth, and low annual evaporation rates. Permafrost limits groundwater movement, and the water budget of the tundra is dominated by three components, precipitation, mostly as snow, evaporation (including evaporation from snow) and runoff (Rydén 1981) (Figures 2a and 2b).

Peat development is most dramatic in the great raised mires of northern Europe. Though spring snowmelt can still be a major input component, annual precipitation is much higher than in the polar regions. Raised mires and blanket mires are isolated from the regional groundwater body, and outflow from the bog is as surface and near-surface runoff (Figures 3a and 3b).

Many lowland wetlands develop as fens in groundwater discharge areas, and groundwater-fed wetlands are particularly important where precipitation is low or strongly seasonal. Interactions between wetlands and groundwater bodies are a common feature of East Anglia, UK, and the Netherlands, where groundwater inputs can exceed those from precipitation. Koerselman (1989) computed a water budget for a small area of

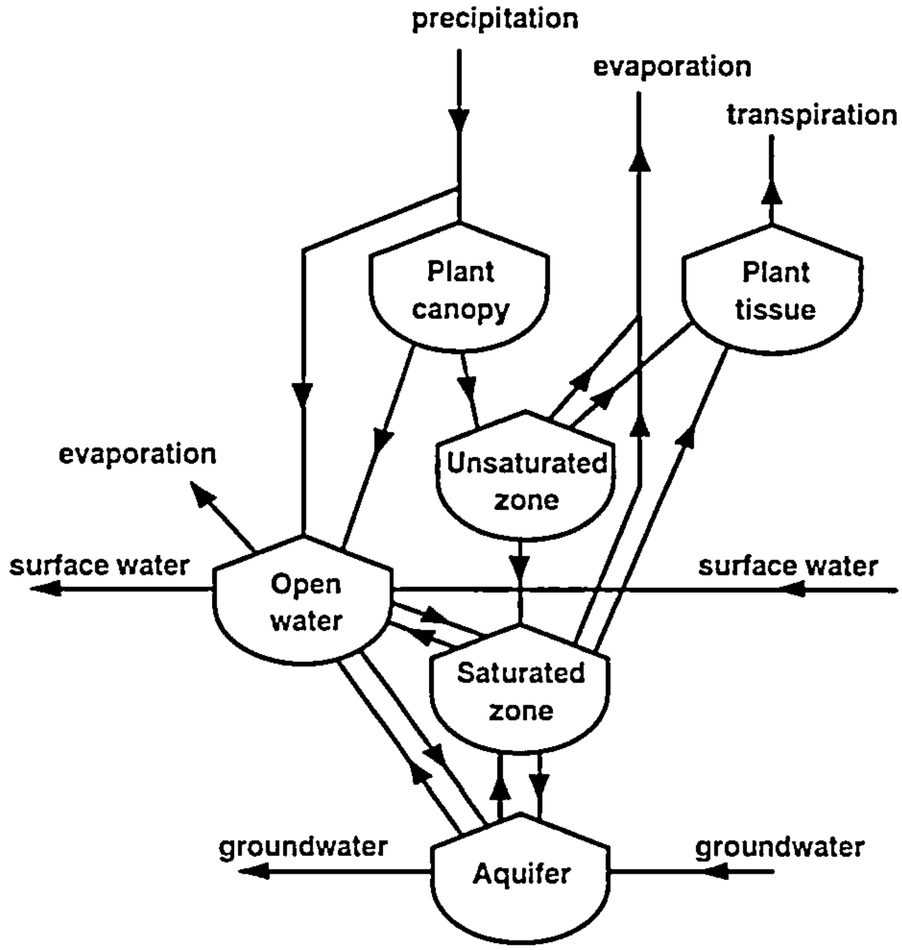

Figure 1 The storage of water in a wetland takes place in permanent open water bodies or as temporary flooding, in the unsaturated zone of the soil (many wetlands, e.g. meadows and grazing marshes, have soils that are merely moist) and in the saturated zone. The wetland groundwater body may be isolated from the regional groundwater body by impeding or confining layers. The flows between these various storages are of vital importance in the water quality functions of wetlands (Patten & Matis 1984).

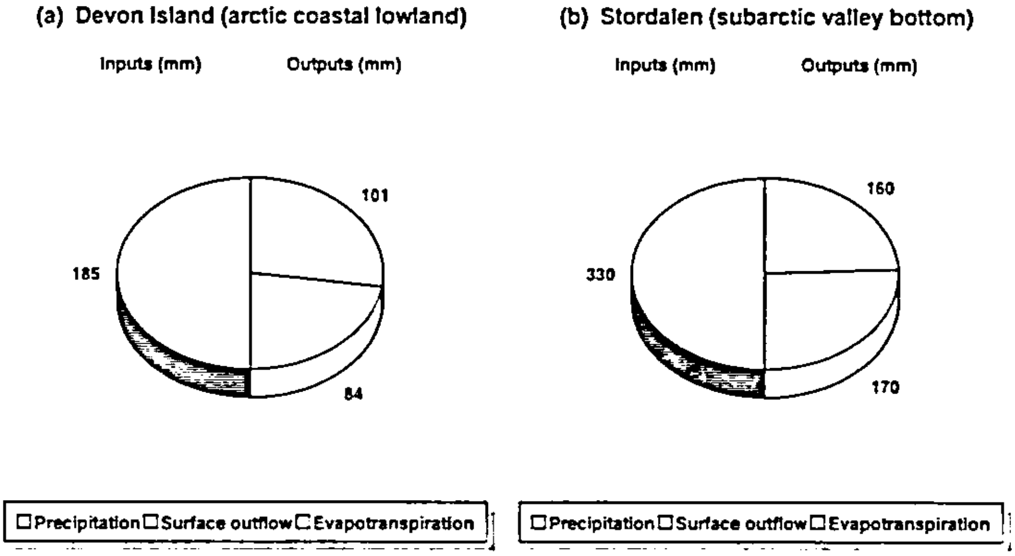

Figure 2 Wetlands in high latitudes are characterised by low precipitation (mostly as snow) and evaporation rates that are limited by insolation and temperature. These water budgets were presented by Rydén (1981), and relate to periods of 3 years and 5 years respectively.

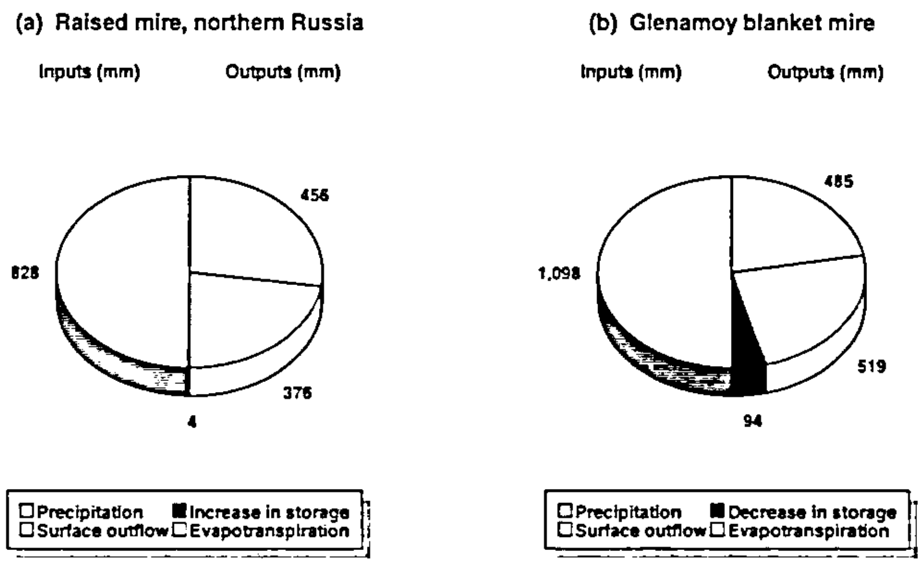

Figure 3 Bavina (1975) presents a water balance for an un-named mire (a) in the north of the forest zone of the European USSR. The components of the water budget are similarly distributed in a blanket mire (b) studied by Burke (1975), near Ireland's west coast, where "snow and frost are almost unknown". Bavina's water budget is an average computed from 1955 to 1966: Burke's relates to an 11-month period in 1968.

fen from which much of the original peat had been harvested, and found that upward seepage from an underlying sand aquifer was an important contributor (Figure 4). An extreme case of groundwater dominance is the spring-fen investigated by Gilvear et al. (1993) at Badley Moor in Norfolk, UK, where localised groundwater upwelling has resulted in the build-up of tufa mounds. During a drought period in 1988-9, groundwater discharge made up 90% of the water inputs and surface water outflow 87% of the output.

The largest wetlands in the world are complex systems driven by river flow. Storage of water is mostly in the form of flooding of enormous areas by overbank flow, and evaporation rates are high. The Okavango Swamp in northern Botswana is an inland delta, across which floodwaters are dissipated. In an average year evaporative losses are about 1.5 m, 50 times the surface water outflow (Dincer et al. 1987).

Notwithstanding the value of the water balance as a research tool, it is instructive to look at how water budget calculations have been used in the pursuit of greater objectives. Development of various kinds is a constant backdrop to the field of wetland science, and even the tundra is not immune to human interference (Radforth 1977). The work of Rydén(1981) was associated with ecological studies, and with climate and water availability under extreme conditions where the scars of development would heal at very slow rates. Bavina (1975) was concerned with drainage and reclamation of bog land, and the potential loss of wetland functions, and found that mires had no effect on stream flow in normal and wet years, while there was some additional loss of water in dry years. Lowland mires could also delay runoff from surrounding high ground. Burke (1975) also concluded that widespread drainage would have beneficial effects on river flow. Conversely, the objectives of Gilvear et al. (1993) and Koerselman (1989) were to produce useful information for the conservation of fens in the face respectively of groundwater abstraction and the introduction of river water into ditches.

In all of these studies the water budget provided essential support for decisions on management, development and conservation of wetlands. However, in view of the reservations expressed by Dooge (1975) and Winter (1978) about the uncertainties in water balance computations, it must be emphasised that the story does not end with the production of a simple budget. For a better basis for decision-making it is necessary to delve deeper into the workings of the hydrological system, and to understand thoroughly the interactions of the wetland with the river network, the atmosphere and the regional groundwater body.

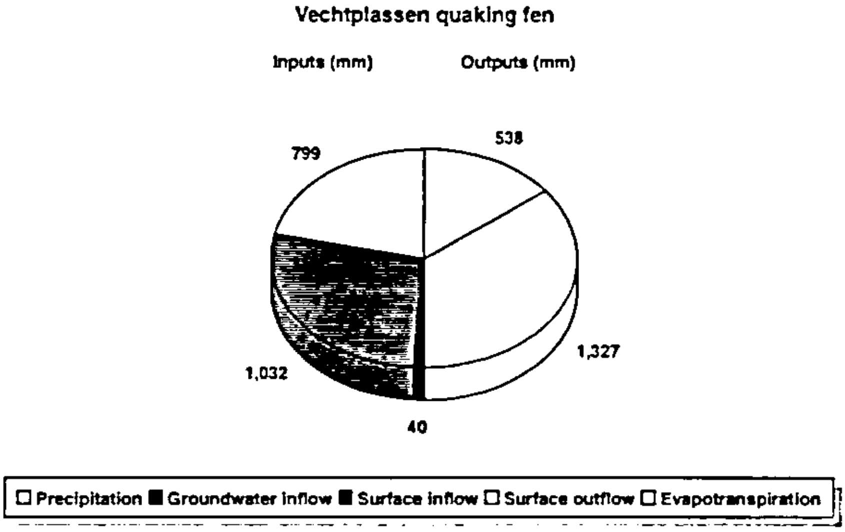

Figure 4 The water budget for a fen (Koerselman 1989) for May 1985 to April 1986 includes significant groundwater and/or upland surface water inputs, which have implications for water quality. Groundwater inputs, being seasonally smoothed, also tend to have a stabilising influence on water levels. Perhaps because of the difficulty of quantifying the groundwater components, there are few well-documented studies of the overall water budget of fens and marshes.

Interactions with rivers - the regulating role of wetlands

Damage and loss of life caused by floodwaters have justified large-scale civil engineering works across the world. Success rates of these works have been variable, but one common feature has been the destruction of wetland habitats and lifestyles, as reclamation and development of floodplain land have followed close on the heels of the engineers. A more flexible approach, conforming to and using natural processes, is now in vogue, and the perceived functions of wetlands are being investigated more closely (Carter et al. 1978; Larson 1982).

The regulating function of wetlands has often been quoted as an argument for preservation: but while extensive inundation is clearly a storage mechanism for large discharges, seasonal regulation is less easy to demonstrate. Indeed evidence suggests that wetland evaporation usually reduces summer base flows in rivers, while winter precipitation encounters a store that is already full, and runs off directly (Vecchioli et al. 1962).

It is often been argued that drainage or channelisation can increase base flows from wetlands (Verry & Boelter 1978), as a result of increased interception of shallow groundwater (Daniel 1981). It is also true that drainage schemes single out groundwater discharge areas within wetlands for special attention, and the proportion of deeper groundwater in outflowing streams increases as a result. In undisturbed wetlands, flat topography and inefficient channels in the natural drainage network together produce significant reductions in flood peaks, and Novitzki (1978) reported 80% lower flood flows from basins in Wisconsin that had lake and wetland storage. Artificial drainage can generate increased base flow from the wetland itself, but it can also increase flood peaks and accelerate the recession (Nicholson et al. 1989), even to the extent of increasing flows throughout the entire range, as shown by the flow duration curve of the Albemarle Canal in North Carolina (Daniel 1981). Upland bogs in their natural state provide some short-term regulation, but summer flows from peat are small (Bay 1968) and what base flow there is is likely to originate from aquifer storage (Verry & Boelter 1975).

The water balance of a wetland gives an overall picture of how the wetland functions when considered as a unit, but there are circumstances under which it is of interest to understand what happens within the wetland. For instance, streams flowing through a wetland area can have varying degrees of interchange of water and dissolved solids with their surroundings (Ming-Ko Woo & Valverde 1981): straightening and deepening of natural channels changes the flooding regime of wetlands, and it has important consequences for water quality, for example in reducing the effectiveness of wetlands for detaining and attenuating nutrient transport into Lake Okeechobee, the principal freshwater storage site in Florida (Ammon et al. 1981).

It is doubtful whether the function of seasonal regulation can ever be carried out efficiently by wetlands, as changes in stored water over the dry season are due to evaporative demand rather than drainage (Verry & Boelter 1978). The function of seasonal regulation belongs instead to deep aquifers, whose reserves are protected from transpiration. Verry & Boelter (1978) pointed out that the uniformity of streamflow from fens could be attributed to a large discharge of groundwater into the wetland. Where evaporation rates are high, regulation of wet season runoff by wetlands is usually at great cost to the annual discharge. For instance the White Nile leaving the Sudd in southern Sudan has a much-damped seasonal hydrograph, at the expense of about half its mean flow (Sutcliffe & Parks 1987).

Interaction with the atmosphere - evapotranspiration from wetlands

Evaporation is a physical process that depends upon atmospheric parameters, physical constants related to water, and climate variables. Evaporation takes place from open water, moist soil and other wet surfaces, notably plant surfaces which have intercepted precipitation or dew. The related process of transpiration, the loss of water from within the aerial parts of living plants, is also defined in part by physical processes, but there are added complications of biological control through stomatal response, tending to reduce the opportunity for transpiration, and other physiological effects such as loss of turgidity (i.e. wilting) or adaptations resulting in enhanced interchange with the atmosphere (e.g. the quivering leaves of aspen).

In hydrological usage, evaporative losses are generally considered as a combined process of evapotranspiration, which is arguably the most complex component of the water balance, the most difficult to measure, and as luck would have it is often the most important. The study of evapotranspiration from wetland sites has a long history and a complexity far beyond the scope of this paper, and only a few illustrative studies will be described here (for thorough and critical literature reviews see Ingram 1983 and Crundwell 1986).

In view of the importance of evaporative losses to water resources, it is perhaps surprising to find that many of the earliest serious attempts to measure wetland evapotranspiration were carried out by academic botanists, followed closely by hydrologists trying to make the best of the meagre water resources of the American West.

The most immediate water resources problem to be addressed was the potential for losses from surface water storage reservoirs due to encroaching floating and emergent plants. It was recognised that enhancement, if any, would be related to climate in the same way as open water evaporation, and most evapotranspiration studies have used as a standard for comparison either open water evaporation E_0 (sometimes estimated from evaporation pan results) or potential evapotranspiration PE_t calculated according to one of a range of empirical or theoretical equations. The use of several different standards has led to a confusing picture, which has left the fundamental question still open to debate. Another complicating factor is the difficulty of carrying out controlled experiments without encountering the "oasis effect" of advected energy from surrounding drier and therefore hotter surfaces and the "clothes-line effect" of increased exposure to drying winds (Linacre 1976). Both have been implicated in over-estimation of wetland evapotranspiration.

The experiments themselves have been mainly of two types: Bowen ratio methods, which assume laterally isotropic and homogeneous conditions (equivalent to an infinite extent of uniform crop), and lysimeter methods, in which it is necessary to assume that conditions in an isolated tank are similar to those in the undisturbed crop and its substrate. Clearly, some experimenters have come closer than others in their approach to these ideals. Occasionally the more sophisticated eddy correlation instruments have been used.

For plants growing in standing water, evapotranspiration is the sum of transpiration and evaporation from the water surface. The evaporation rate would be expected to be reduced below open water values by sheltering from sun and wind, reduction in surface area, inhibition of wave action etc., and the question is whether transpiration fails to make up this deficit or exceeds it. Floating-leaved plants such as water lilies and water ferns would be expected to have a small inhibiting effect on evapotranspiration, and results presented by Anderson & Idso (1985) confirmed that water fern reduced evaporation by 10% for a 95-100% cover, while water lilies reduced evaporation by 15% for 100% cover. Water hyacinth (*Eichhornia crassipes*) has a canopy which rises above the water surface as growth proceeds, and evidence presented by van der Weert & Kamerling (1974), Anderson & Idso (1985) and Snyder & Boyd (1987) demonstrated that evapotranspiration exceeded open water evaporation by 44-48%, 50% and 75% respectively, though van der Weert & Kamerling went on to suggest that the ratio depended upon climatic factors, particularly vapour pressure deficit.

In favourable conditions, emergent plants colonising open water margins can form extensive stands and floating mats, usually dominated by a single species. Familiar and almost ubiquitous emergents are the reedmace (cattail in the US), *Typha spp.*, and the common reed, *Phragmites australis* (*Phragmites communis* in early literature). Perhaps because of the water resources implications, research work on reeds and similar plants has concentrated on swamp conditions, with standing water, rather than the drier fens in which certain of these plants are also dominant.

The tens of thousands of small lakes in the prairies of North America have long attracted the attention of hydrologists. Eisenlohr (1966a & 1966b) found that prairie potholes with a cover of bulrush (*Scirpus = Schoenoplectus spp.*), reedmace (*Typha*) and similar emergent plants had lower evapotranspiration rates than potholes with open water. Investigation of seasonal variations showed that this result reflected the short growing season: outside of this season evaporation was reduced considerably by sheltering by dead stems, but this effect was outstripped by transpiration during the late summer. This draws attention to an important aspect of wetland evapotranspiration studies: the effects of climate should not be taken lightly, and care must be exercised in the extrapolation of results from one location to another, even allowing for overall differences in annual potential evapotranspiration or open water evaporation.

A similar but more serious problem is the loss of streamflow brought about by phreatophytes (mainly trees) growing in the riparian strip alongside streams in semi-arid zones (Meyboom 1964): at their most extreme, a community dominated by the western cotton-wood, *Populus sargentii*, the losses per unit area from these riparian wetlands were estimated to be up to 2.5 times those from open water. It is perhaps because of the "oasis effect" that these evaporation rates are far in excess of those found by Priban & Ondok (1986) for willow carr in the moister conditions of the Trebon Biosphere Reserve in Czechoslovakia.

The effects of dead material in the standing vegetation were noted by Rijks (1969) in a study of evapotranspiration from an old papyrus stand in Uganda. Papyrus is renewed by fire, and evapotranspiration of the new crop proceeds at about the open water rate (Penman 1963). In contrast, Rijks found that evapotranspiration from the 6-year old stand was 60 ± 15% of E_0, and attributed this to sheltering from wind and conversion of incoming solar radiation into sensible heat at the level of the papyrus heads.

Eddy-correlation techniques, corroborated by Bowen ratio measurements, were used by Linacre et al. (1970) to determine the summer rates of evapotranspiration from a *Typha* swamp in Australia. There was considerable scatter between results for individual days, but the general conclusion was again that the growth of emergents reduced evaporation to about 70% of E_0. Drawing attention to the change in the ratio after rain had "freshened the reeds and enhanced evapotranspiration", Linacre et al. (1970) suggested that reduced growing season evapotranspiration might be a feature of wetlands in dry climates.

Growth of common reed, *Phragmites australis*, in the lakes and fishponds of Central Europe is vigorous, and summer transpiration rates could be expected to be high. Smid (1975) carried out Bowen ratio measurements on several days between June and October 1973, and found that evapotranspiration exceeded open water evaporation (measured by a floating pan) by 85% at the end of June, with approximate equality at the start of June and the start of October.

The extrapolation or regionalisation of the results on intensive measurements always presents a problem. Bernatowicz et al. (1976) attempted to make their work more applicable by expressing their results in terms of a transpiration coefficient, the amount of water transpired in the course of the production of unit dry weight of the above-ground crop. Armed with a reliable estimate of this coefficient for a given species, the hydrologist could estimate the total transpiration of a water body or wetland in any given season. Bernatowicz et al. measured the coefficient for four species, and found values ranging from 320 for reed (*Phragmites australis*) to 593 for bulrush (*Schoenoplectus lacustris*). Estimates of evapotranspiration from four Polish lakes, allowing for evaporation from the standing water, were between 90 and 130% of E_0, depending upon the above-ground dry weight, which ranged from 5.4 tonnes/ha to 10.0 tonnes/ha.

There are many variations on the lysimetric method for determining evapotranspiration, and the diurnal fluctuation method could be considered a lysimeter variant (Godwin 1931; Heikurainen 1963; Dolan et al. 1984; Laine 1984; Gilman 1993; Gilman 1994). In wetlands with a water table close to but below the ground surface, evapotranspiration causes the water table to fall steeply during the daylight hours, and the water table then rises or falls more slowly during the night as other processes, without such a strong dependence on solar radiation, take over.

Using this method Heikurainen (1963) and Laine (1984) explored the effects of tree species and water table elevation on evapotranspiration from forested bogs. Transpiration was close to the potential evapotranspiration rate, and there were only slight differences between plots carrying different vegetation communities. Laine concluded that differences in water yield from different communities were due to differential interception of precipitation. Interception, in this case by dwarf shrubs such as heath, may also have been implicated in Eggelsmann's (1963) observation that evapotranspiration rates from an uncultivated raised mire exceeded the potential rate in winter, despite summer evapotranspiration rates of about 60% of the potential rate. However, measured bog evapotranspiration rates appear to have a wide spread: Sturges (1968) reported summer evapotranspiration from a spikegrass (*Eleocharis pauciflora*) community at about 80% above the potential rate, and Nichols & Brown (1980) measured evaporation rates from Sphagnum moss at between 130 and 150% of the open water evaporation rate.

The lakeshore marsh in Florida instrumented by Dolan et al. (1984) carried a mixed emergent community, and the water table was on average 0.2 m below the soil surface. It is not easy to maintain daily measurements throughout the year, and this study went on to explore the possibility of using a simple evapotranspiration model based on the atmospheric saturation deficit and the above-ground live biomass to arrive at monthly and annual totals. Indicative results, based on an off-site evaporation pan, suggested a seasonal trend, with wetland evapotranspiration exceeding open water evaporation during the growing season by up to 65% (September and October), but falling to about half of E_0 in the winter (February and March).

Fens have received little attention in the literature. The study by Koerselman & Beltman (1988) presents a plausible model of the hydrological behaviour of sedges in floating fens in the Netherlands. A lysimeter study showed that in low-growing sedge-dominated communities the evapotranspiration was between 74% and 81% of E_0 (i.e. approximately the same as the potential evapotranspiration. Koerselman & Beltman mentioned interception of rainfall as a complicating factor, and suggested that vegetation structure, in the form of leaf inclination, might be an important controlling variable. Confirmation of the appropriateness of the potential estimate for low-growing sedge comes from a study carried out near James Bay, northern Ontario (Lafleur 1990). At a wet site, which started the three-month growing season with standing water, growth of vegetation reduced evapotranspiration rates to about 85% of open water evaporation, while a drier site always evaporated at about 60% of E_0. Rather higher evapotranspiration rates were measured by Roulet & Ming-Ko Woo (1986), also in northern Canada, where evapotranspiration from a sedge fen was found to equal evaporation from a nearby lake.

The pattern for more recent, and probably for future, wetland evapotranspiration studies appears to be the exploration of the detailed relationships between the transpiration process, the inhibition of evaporation where standing water is present, and the structure of wetland vegetation (Ingram 1983). There is a need for uniformity of approach, to encourage comparison of results and the development of genuinely transferable models. In the interests of the wise use of wetlands, it is important that the results of scientific investigations be used intelligently, and it is to be hoped that the models of wetland evapotranspiration that emerge will be sufficiently simple and powerful to see wide application.

Interaction with the regional groundwater body

Groundwater is frequently a contributor to the input side of the wetland water balance. It is not generally appreciated that wetlands, by detaining water in its rush to the sea, can also recharge the groundwater resource, and that some wetlands have a complex interaction with the groundwater body, with different areas of the same wetland acting as groundwater recharge and discharge areas (Schot, Barendregt et al. 1988; Gehrels & Mulamoottil 1990), and the wetland area acting as recharge or discharge zone at different seasons (O'Brien 1977).

The largest body of evidence for complex interactions comes from work done on prairie potholes, though there must be many other areas, for instance the Breckland Meres of Norfolk (Denny 1993), that would justify detailed investigation. Outward seepage from marshy prairie ponds investigated by Eisenlohr (1975) acted as a source of recharge to groundwater mounds. The majority of water flowing radially outwards from the pond was used locally by plants rooting into the water table, but outward seepage was good for the pond, as it helped to regulate the build-up of dissolved solids, mainly sodium sulphate. A more detailed and extensive study by LaBaugh et al. (1987) presented a picture of ponds in the same general area acting as sources, conduits and sinks of groundwater. Again water chemistry was important in understanding the setting of each wetland in relation to the groundwater flow system, and conversely, the key to accounting for temporal and spatial variations in chemistry lay in the hydrology of the wetlands.

Many wetlands in river valleys are strongly influenced by river stage, and the transfer of water between wetland and river at intermediate stages is by groundwater flow through alluvial sediments. River management usually acts to lower both flood and low flow stages, leading to a decline in water levels and water supply to the wetland. Where wetlands are important for conservation, there is consequent decline and loss of habitat. Concern over the habitat of wildfowl including migrating cranes led to a study of the hydrogeology of floodplain wetlands alongside the Platte River in Nebraska (Hurr 1983), where groundwater flow is sustained by the river stage, and subject to threat both from abstraction of river water, and of groundwater from the alluvial aquifer some distance from the river. The case highlights the importance of understanding the connections between wetland groundwater and more extensive groundwater bodies exploited for water supply.

Lakes and wetlands are often viewed as having a constant relationship with their surroundings, perhaps modified by seasonal changes, but essentially stable. This simplistic view was questioned by Winter (1983), who used a two-dimensional saturated/unsaturated groundwater model to demonstrate that the distribution of recharge was spatially non-uniform, and groundwater mounds developing beneath depressions could create "groundwater dams" which controlled the direction and velocity of groundwater movement, including seepage through lake beds and wetland substrates. This concept of non-uniformity of groundwater behaviour brought about by flow processes, added to the non-uniformity of hydraulic properties, introduces a whole new range of implications for groundwater and surface water quality. The theme of inherent complexity was carried forward in a review paper by Carter & Novitzki (1988), who used a small number of intensively researched case studies in Indiana, North Dakota and Virginia/North Carolina to demonstrate "the impossibility of designating all wetlands either discharge or recharge areas".

The approach to this problem put forward by Siegel (1988), which he designates "the hydrogeologic systems approach" is the construction of a conceptual model incorporating a complete description of the geological framework, including hydraulic properties of the various media, with groundwater flows characterised by partial differential equations. Siegel is critical of researchers who have sought simplification in assumptions such as hydraulic isolation of peat from its mineral substrate. As an example he quotes the case of a raised bog in the Lost River Peatland, Minnesota, where ombrotrophic conditions occur at the surface of a peat mound sustained from below by groundwater from mineral horizons. The recharge area for this flow may be a beach ridge some 12 km away. As a first effort towards understanding the transient behaviour of the recharge/discharge function, Siegel recommends the installation of piezometers in the organic and mineral soils of wetlands rather than dipwells in the upper horizon only.

Overview and conclusions

So diverse are wetland systems and the accidents of climate, topography and geology that have created them that it could be said that the only common feature was the dominance of hydrological processes as driving and controlling mechanisms. As the central concept of hydrology, the water balance has a unique position in scientific wetland management, conservation and restoration.

The contribution of the simple water balance to our understanding of wetland hydrology is that it expresses the relative importance of the various water sources and hydrological processes, but it is only when the detail of the water balance is filled out, both spatially and temporally, that it can become a sound foundation on which to base wide-reaching decisions, for example to resolve conflicts between development and conservation. We need to know the functions of the different parts of a wetland area, the seasonal variation in water balance components and the effects of drought years and time-varying man-induced impacts.

The most urgent challenges in wetland hydrology lie in the interactions of the wetland with its surroundings. After direct reclamation, which has been slowed down to an extent by international pressure for conservation, the greatest threats to wetlands come from activities such as intensive agriculture and groundwater abstraction on adjacent land. The value of wetland sites in terms of their hydrological functions, particularly relationships with river flow and quality, is beginning to be widely appreciated, and attempts are being made to quantify wetland functions so that planning decisions can be put on a firmer footing.

It is a tradition that every scientific paper ends with a call for further research. In the case of wetland hydrology, that call comes not just from the researchers, but also from decision-makers and planners who recognise the need for wise and sustainable development, and from conservationists who are aware that arguments for preserving wetland habitat could be strengthened by a clearer appreciation of the influence of wetlands on the water environment, and of their potential role in environmental management.

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BROADLEAVED WOODLAND, ITS WATER USE AND SOME ASPECTS OF WATER QUALITY

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Abstract

The effects of broadleaved woodland on water quantity and quality are considered by examining the results of a detailed hydrological study of ash (*Fraxinus excelsior*) and beech (*Fagus sylvatica*) woodland in southern England in the wider context of other European studies. Estimates of the water use of broadleaved trees growing on two soil types are presented using a simple model requiring only daily estimates of precipitation and potential evaporation. With some important exceptions the transpiration from most species of broadleaves grown in northern Europe is conservative. The major uncertainty in estimating the water use of broadleaved woodland is due to the wide variability in interception loss.

Introduction

As a result of EU common agricultural policy there is in the UK a continuing policy to take farmland out of food production and this is recognised as an opportunity, through conversion of farmland to woodland, to make the UK less dependent upon imported timber. The Woodland Grant and Farm Woodlands Premium schemes offer financial incentives to landowners and farmers to grow trees, especially broadleaves. In contrast to the major forestry effort to date, the new plantations will be concentrated in the drier, lowland regions of the UK. With the additional incentives of the Community Forest Programme, which envisages 12 community forests in England each covering 8000 to 20000 ha and the recently approved new National Forest (over a proposed area of 500 km^2 in the Midlands) it is likely that the area of conventional broadleaved woodland will continue to increase in lowland Britain. There is also increasing interest in the growth of short-rotation coppiced poplar and willow as a supply of non-fossil fuel; there are several demonstration schemes on farms in the south of England.

The latest figures for the two grant schemes (R. Kerr, Forestry Commission, pers. commun., 1994; N. O'Brien, MAFF, pers. commun., 1994) indicate that the largest percentage of the grants issued are for plantations of mixed broadleaved species followed by mixes of native broadleaved and broadleaves and conifers. The ranking of the single species plantations are: birch (*Betula* sp.) predominantly in Scotland, oak (*Quercus* sp.), ash (*Fraxinus excelsior*), beech (*Fagus sylvatica*), gean (*Prunus avium*) and sycamore (*Acer pseudoplatanus*).

In the past coniferous afforestation of the high rainfall upland areas has been associated with wide ranging and occasionally adverse effects on regional hydrology, hydrochemistry and ecology. In particular it has been shown that evaporation of water from coniferous plantations can considerably exceed that from adjacent grassland areas, resulting in reduced stream flow. Any reduction in water supply in southern England, possibly arising from the new broadleaved woodlands, would have serious consequences because of the small difference between rainfall and evaporation.

In a review of 94 catchment studies worldwide Bosch and Hewlett (1982) concluded that on average a 10% change in coniferous forest cover results in a 40 mm a year change in water yield whereas the same change in deciduous forest cover results on average in a 25 mm annual change in water yield. The reason for the increased water use after afforestation is increased interception loss, precipitation which is intercepted by the vegetal canopy (and litter layer) and then evaporated directly back into the atmosphere having never reached the soil. Sometimes there is also an increase in transpiration after afforestation.

There is also concern about the increasing extent of pollution of surface and groundwaters from fertilizers, pesticides, and various forms of domestic and industrial waste. In the UK the designation of Nitrate Sensitive Areas (NSAs) around existing public supply boreholes in sensitive catchments such as the chalk is now being actively pursued. These NSAs aim to minimize the amount of nitrate leaching by following low-N input strategies. In addition, a national scheme for groundwater protection zones is in force limiting the types of land use allowed adjacent to boreholes. Establishment of woodland could be a suitable option for these sensitive areas.

To predict accurately the impacts of broadfeaf plantation on water quantity *and* quality requires an accurate understanding of the mechanisms controlling evaporation from broadleaved trees. This evaporation occurs through two processes, transpiration and interception loss. The information available for broadleaves on these, and on the controlling mechanisms, are briefly reviewed. The major findings from a recent study carried out in Hampshire and Northamptonshire by the Institute of Hydrology on two species, ash (*Fraxinus excelsior*) and beech (*Fagus sylvatica*) are also reported.

Water use

TRANSPIRATION

Transpiration rates from trees are often less than from agricultural crops because of lower stomatal conductances Jones (1983). According to Roberts (1983), who reviewed published work on forest (conifers and broadleaves) evaporation, transpiration from European forests is a remarkably conservative process with the mean annual transpiration loss being 330 ± 25 mm a year. Moreover the annual transpiration totals are less than the Penman potential evaporation. Roberts suggested the reasons for this were (i) the negative feedback between atmospheric humidity and stomatal conductance and (ii) the generally small effect that variations in soil water content have on transpiration rates. Much research is continuing on the response of stomata to environmental variables but there is much evidence that for many species they respond to a drying atmosphere by closing their stomata thereby reducing the rate of transpiration. Also because of their aerodynamic roughness evaporation from trees is more closely coupled to the atmospheric humidity and windspeed than to radiation, whereas the opposite is true for shorter plants. A change in the stomatal conductance in response to e.g. a change in the atmospheric humidity deficit has a larger effect on the transpiration rate than it does on shrubs or agricultural crops.

There have been a large number of plant physiological studies of broadleaved trees but these have mostly concerned potted or isolated plants, and not all of the species of interest have been studied. There have been many fewer studies of the water use of broadleaved trees in woodlands. Iindeed until a recent study by the Institute of Hydrology (Harding et al., 1992) annual transpiration totals from broadleaved woodland were not available for the UK, but there were several reports for continental Europe. Table 1 presents the available data for broadleaved forests in Europe and shows the notable similarity in annual transpiration totals. The similarity of annual transpiration from conifers and broadleaves in northern Europe implies that broadleaves must transpire at a greater daily rate when the leaves are present than the conifers .

Table 1 Annual transpiration totals for a range of European forest studies

Species	Rainfall (mm)	Transpiration (mm)	Location	Authors
Ash	789	407	Hampshire, England	Roberts & Rosier (1994)
Beech	789	393	Hampshire, England	Roberts & Rosier (1994)
Ash	589	348	Northampton, England	Harding et al (1992)
Oak, Ash, Lime	513	285	European, USSR	Molchanov (1971)
Scots pine	595	353	Thetford, UK	Gash & Stewart (1977)
Norway Spruce	2760	330	Wales	Calder (1977)
Red oak	663	304	Germany	Brechtel (1976)
Oak	663	327	Germany	Brechtel (1976)
Beech	663	301	Germany	Brechtel (1976)
Spruce	827	324	Germany	Ernstberger & Sokollek (1983)
Beech	896	319	Germany	Benecke & van der Ploeg (1975)
Mixed oak	966	343	Belgium	Schnock (1971)

INTERCEPTION

Interception loss from trees is calculated as the difference between precipitation measured in the open, sometimes above the tree canopy, and net rainfall which is usually measured as the sum of throughfall, water with either falls straight through the canopy or drips from leaves and branches, and stemflow. Throughfall is measured using appropriate collectors placed beneath the canopy and stemflow is invariably measured by collecting the water running down the tree trunks using pliable guttering spiralled around and suitably sealed to them. Sampling of both must be thorough to minimise the variability, especially for small storms (Lloyd and Marques, 1987). Because of inadequate sampling many studies of interception loss must be treated with caution.

The interception loss from trees is affected by the combined influences of certain aspects of the plant structure and also the climatic, especially rainfall, regime of the locality where the trees are growing. The interaction of these effects is complicated for deciduous trees because of the variation in leaf cover and seasonality of the rainfall. The main plant parameters affecting interception loss are: the free throughfall coefficient, p (dimensionless), the canopy storage capacity S (mm) and the aerodynamic resistance r_a (sm^{-1}) which for a given set of weather conditions determines the rate at which water evaporates from the surface of wet vegetation. (Because trees are aerodynamically rough they produce efficient mixing of the air above them and r_a is relatively small; for conifers values are typically 3 to 10 sm^{-1} compared to typically 40-50 sm^{-1} for agricultural crops.) Interception loss is therefore dependent upon those environmental factors which affect the values of these parameters. For young plantations the interception loss is likely to be most dependent upon the age of the trees and less dependent upon species. During this time the canopy expands and extends and the amount of rain reaching the ground without interception decreases while the canopy storage capacity increases.

According to Hall and Roberts (1990) the relative canopy capacities (S = (1 ± 0.3) mm, n = 9, compared with S = (1.4 ± 0.6) mm for conifers from 20 studies of 8 species) and values of the bulk momentum-transfer coefficient for broadleaves and conifers suggest that interception loss during the foliated period from some broadleaved species will equal that from some conifers. The results of Leyton et al (1967) and Rogerson and Byrnes (1968) support this showing very similar interception loss during summer from spruce (*Picea*) and hornbeam (*Carpinus betulus*), and red pine (*Pinus resinosa*) and oak respectively.

The results of European interception studies on deciduous tree species made during the last 40 years are summarised in Table 2. The annual percentage interception loss from these studies has been plotted against annual rainfall (Figure 1). As expected for results from a range of species, ages, planting densities and climatic regimes there is much scatter and errors of probably about ±10% can typically be associated with each point. Nevertheless, all of the points lie below the curve fitted to interception data for coniferous forest given by Calder (1982). Further evidence that, on an annual basis, interception loss from conifers exceeds that from broadleaved species is given by various comparative studies made at the same locations e.g. Leyton et al (1967) and Rogerson and Byrnes (1968).

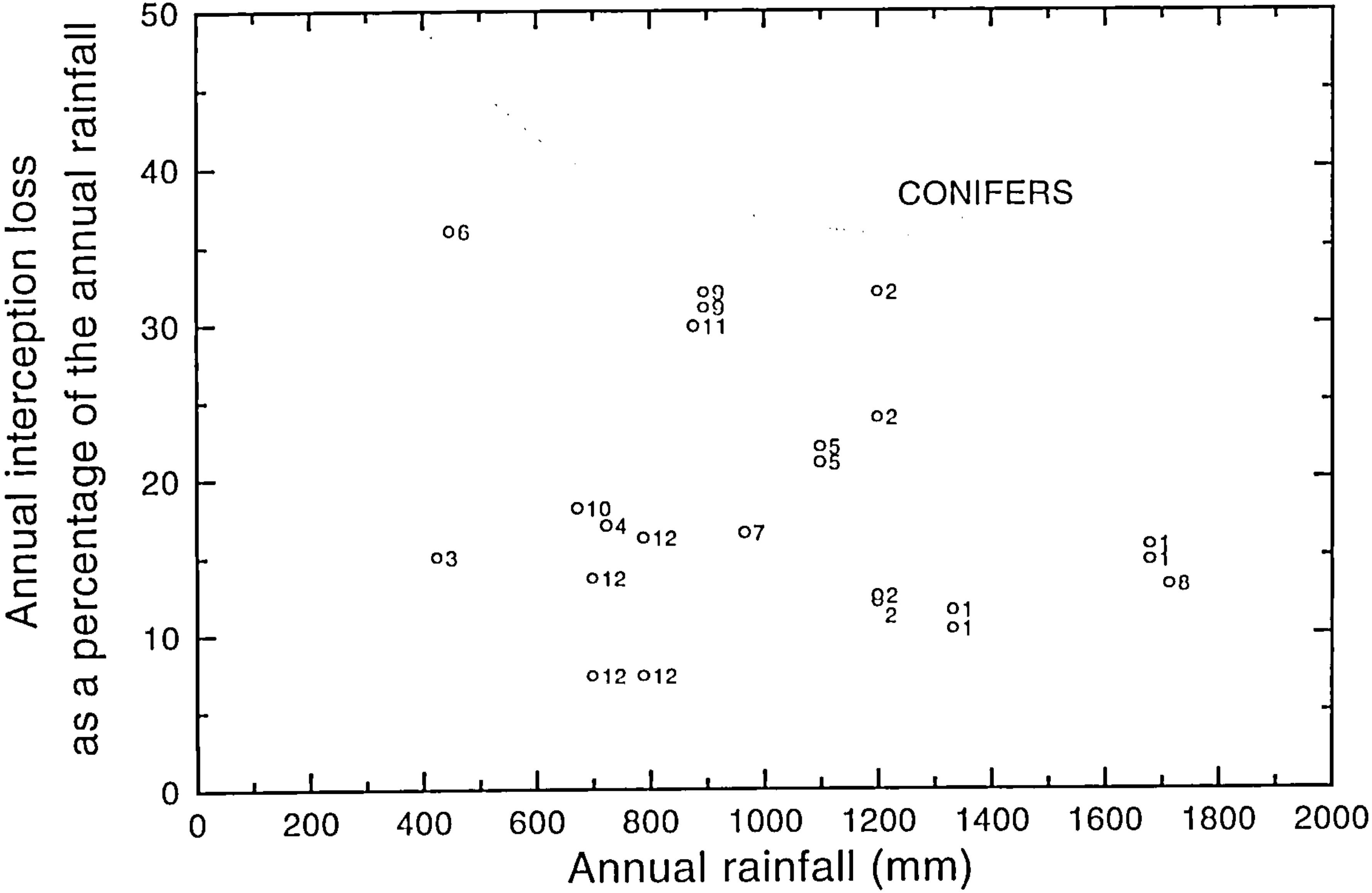

Figure 1. The annual interception loss expressed as a percentage of the annual precipitation plotted against the annual precipitation for European broadleaved trees. The grey line represents the interception percentage for conifers taken from Calder (1982) and the data labels are the reference numbers given in Table 2.

Figure 1. The annual interception loss expressed as a percentage of the annual

White and Carlisle (1967) provided useful comparative values of interception loss for several broadleaved species. They worked in an mixed-aged, multistoried dense (LAI = 8) woodland of coppice mixed with single-stemmed trees covering about 22 ha in Cumbria where the annual rainfall was 1200 mm. Over two leafy and one leafless seasons they measured the annual

Table 2 Annual and seasonal interception loss as a percentage of gross rainfall for European broadleaved trees

Species	Annual Precipitation (mm)	Percentage interception loss Annual	Growing season	Dormant season	Reference	Reference no used in Fig. 1
Alder	1333	11.5			Cape pers. comm.	1
	1680	15.7			Cape pers. comm.	1
Ash	1200	12			White & Carlisle (1967)	2
Beech	425	15	21	6	Aussenac & Boulangeat (1980)	3
Beech/hornbeam	724	17	18.6	15.1	Aussenac (1968)	4
Birch	1099	21			Skeffington pers. comm.	5
Hornbeam	447	36	45	29	Leyton et al (1967)	6
Hornbeam/oak	966	16.5	22.3	10.5	Schnock (1971)	7
Lime	1200	21			White & Carlisle (1967)	2
Oak	1714	13.1	16.9	9.9	Carlisle et al (1965)	8
Oak	895	32			*Ovington (1954)	9
Oak	1333	10.3			Cape pers. comm.	1
	1680	14.7			Cape pers. comm.	1
Oak	1099	22			Skeffington pers. comm.	5
Oak			31	11	Dolman (1987)	
Oak coppice	673	18.1	24	12	Thompson (1972)	10
Oak/birch	877	29.8	23	36	Noirfalise (1959)	11
Southern beech	895	31			*Ovington (1954)	9
Sycamore	1200	23.9			White & Carlisle (1967)	2
Mixed	1200	12.4	16.7	12.1	White & Carlisle (1967)	2
Ash	699	7.3	8.7	7	Harding et al (1992)	12
	789	7.3	7.9	7	"	12
Beech	699	13.6	25.9	11.3	"	12
	789	16.2	23.6	12.3	"	12

*The values given by Ovington were calculated for a three year period after he had established relationships between interception loss and rainfall intensity and total from measurements made over several days throughout a year.

interception loss as 12.4% of the annual precipitation. Additionally they measured the interception loss for individual crowns of different species. The values they obtained are included in Table 2. The dense canopied and large leaved sycamore and lime (*Tilia*) evaporated significantly more intercepted water than either ash or oak which at 13.1% was identical to the figure for oak measured by Carlisle et al (1965) at a high rainfall (1714 mm) site also in Cumbria.

Major findings of a recent hydrological study of ash and beech in southern England

In the light of the paucity of water use data for preferred broadleaved tree species in the UK the Department of Environment and National Rivers Authority funded a research project Harding et al. (1992) to examine the effects of broadleaved woodland on water quantity and quality in lowland England.

Ash and beech were studied at two locations: Black Wood, a 2.7 km^2 Forestry Commission woodland on ~25 cm dark brown silty loam overlying the Upper Chalk in Hampshire [51° 10' N, 1° 15¢ W]; Old Pond Close a 25 ha Forestry Commission wood at the eastern end of Yardley Chase in Buckinghamshire, on dark greyish-brown slightly calcareous clay-loam overlying Oxford Clay [52° 10¢ N, 0° 42¢ W]. Black Wood is primarily mature beech but with a small plantation of low grade ash within it. Old Pond Close is predominately mature ash with a small number of oak and with a vigorous understorey. Details of the sites and measurements are given in Harding et al (1992), Neal et al (1991) and Roberts and Rosier (1994).

The water use of the beech and ash, including the understorey, at Black Wood was studied between spring 1989 and summer 1991. Measurements were made of: soil water content and potential, the weather above the ash and beech canopies and beneath the ash canopy using automatic weather stations, net rainfall beneath both species and gross precipitation. The water quality was also studied. Samples of rainfall, stemflow and throughfall and groundwater samples from 21 boreholes were collected for chemical analysis. Measurements were made at Old Pond Close from April 1990 to August 1992 and included: soil water content and potential, the weather above and below the ash canopy, net and gross precipitation. At both sites measurements were made of the leaf area index and, on selected days, stomatal conductance (g_s), and leaf water potential (y_1). It is not appropriate here to describe these measurements fully but proven techniques were used and details can be found in the references cited above.

ESTIMATION OF WATER USE: TRANSPIRATION

The transpiration for both tree species at both sites, and for the understorey in the ash plantation at Black Wood, was estimated using a multilayer model (Roberts and Rosier, 1994). In this model the total transpiration λT is taken as the sum of the transpiration from each of m canopy layers where measurements were made estimated using the Penman-Monteith formula (Monteith, 1965) viz.

$$\lambda T = \sum_{i=1}^{m} \lambda T_i = \sum_{i=1}^{m} \frac{\Delta' R_{n,i} + \rho c_p \delta q g_{a,i}}{\Delta' + c_p / \lambda (1 + g_{a,i} / g_{c,i})} \tag{1}$$

where: $R_{n,i}$ is net radiation (assumed equal to the available energy) for the ith layer, c_p is the specific heat of air at constant pressure, $g_{a,i}$ is the boundary-layer conductance of the ith layer, $g_{c,i}$ is the canopy conductance of the ith layer calculated from $g_{c,i} = L_i^{*} g_s$ where L_i^{*} is the leaf area index of the ith canopy layer and g_s is the mean stomatal conductance of the leaves in layer i, δq is the above-canopy specific humidity deficit of the air, Δ' is rate of change of saturated specific humidity with air temperature, λ is the latent heat of vaporisation of water and ρ is the density of air.

Roberts and Rosier (1994) describe in detail how equation (1) was used with measured surface conductances to estimate hourly transpiration for both species at Black Wood for the period of the experiment (June 1989 to November 1991). Briefly, a three-layer version of the model with weather data from above the canopy was used to estimate the beech transpiration for the leafed period only. For the ash (including the understorey) a four-layer version of (1) was used with weather data from above and below the canopy for all months when vegetation was present.

For 1990 and 1991 Roberts and Rosier calculated annual transpiration totals from ash of 440 mm and 336 mm respectively and for the same years 403 mm and 334 mm respectively for the beech (Table 1). Although the beech transpires more each day during the summer months the transpiration from the ash understorey over a longer period more than compensates. On average the understorey contributed about 46% of the annual transpiration from the ash plantation. There is obviously a strong seasonal variation in this percentage which ranges from 20-30% during the foliated period (May to September) to 100% in other months.

Daily transpiration totals T_D derived from the hourly values are plotted against daily Penman potential evaporation (E_T) values (also calculated from the above-canopy weather data) in Figures 2 and 3. In doing this we are using E_T purely as an index of the evaporative demand of the environment. The graphs show that the ash daily transpiration did not achieve E_T and the beech transpiration rate was less than E_T at the higher values. It is likely that this is the effect of reduced stomatal conductance at high values of specific humidity deficit.

To make it possible to estimate daily transpiration (T_D) from beech and ash using the readily available daily values of E_T, Harding et al. (1992) derived a simple model using the daily transpiration estimates. For beech a simple polynomial was fitted for the leafed period (days 150 to 270), Figure 2:

$$T_D = 1.63\, E_T - 0.15\, E_T^2 - 0.41 \tag{2}$$

with transpiration put to zero in the unleafed period.

For the ash a simple proportionality was used for the leafed period viz.

$$T_D = 0.74\, E_T \tag{3}$$

shown in Figure 3. However outside the leafed period an understorey component was estimated using

$$T_D = 0.1\ E_T \tag{4}$$

for before day number 90 and after day number 270. Between day numbers 90 and 150, ash transpiration was estimated by

$$T_D = (0.64\,(i - 90)/60 + 0.1)\, E_T \tag{5}$$

where i is the day number.

ESTIMATION OF WATER USE: INTERCEPTION

A detailed analysis using an interception model of the data collected for all seasons at the Black Wood and Old Pond Close provided information on the interception characteristics of both the foliated and unfoliated canopies of beech and ash. Values of p, S and r_a of 0.4, 0.43 mm and 10 sm^{-1} (at a windspeed of 1 ms^{-1}) for ash and 0.25, 0.92 mm and 12.5 sm^{-1} (at a windspeed of 1 ms^{-1}) for beech respectively. These figures indicate that, unlike some other broadleaves (Leyton

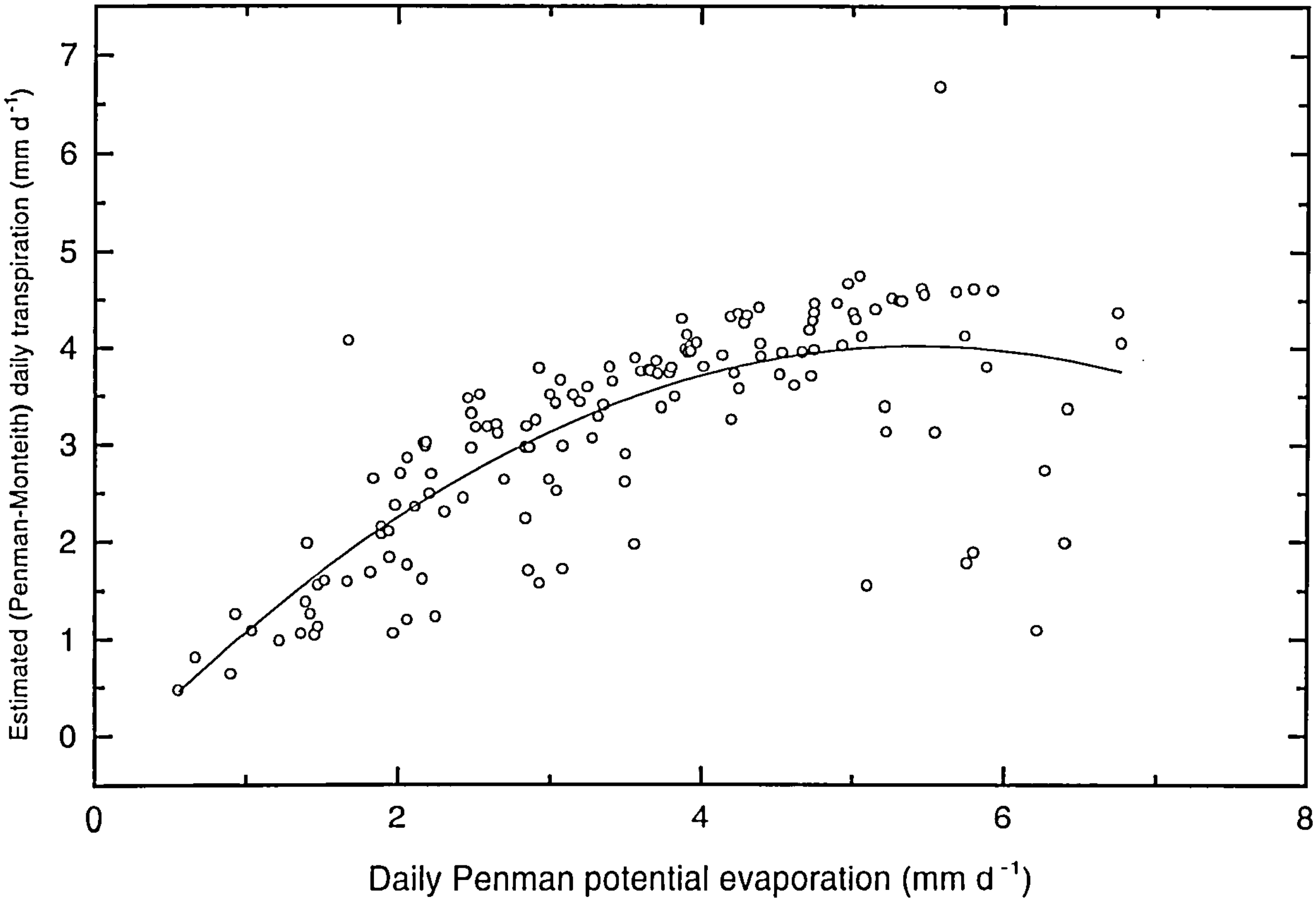

Figure 2. Daily transpiration plotted against E_T for beech at Black Wood, 1989-1991, with the fitted curve $T_D = 1.63 E_T - 0.15 E_T^2 - 0.41$.

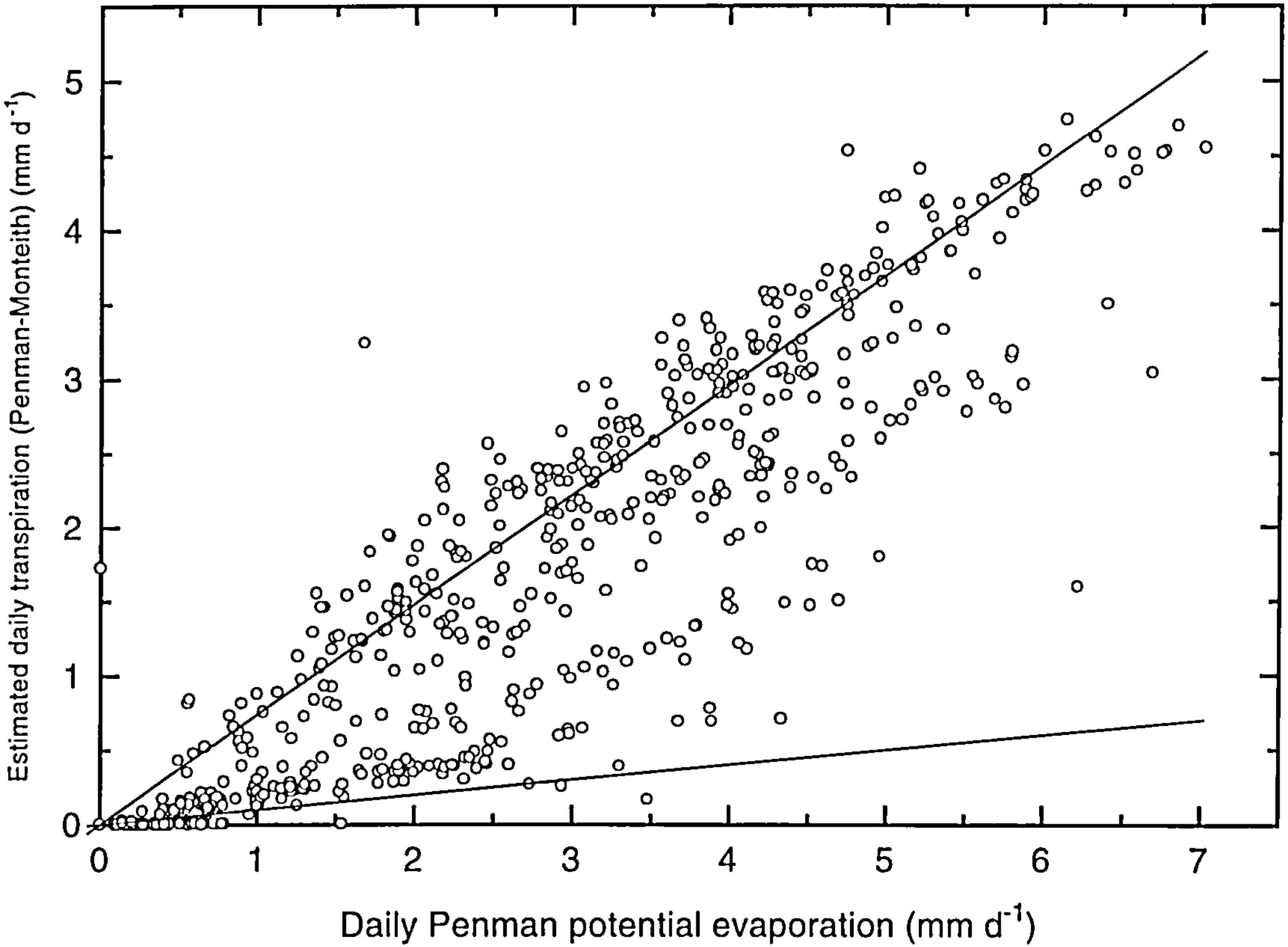

Figure 3. Daily transpiration plotted against E_T for ash at Black Wood, 1989-1991, with the fitted lines $T=0.74E_T$ and $T=0.1E_T$

et al, 1967; Rogerson and Byrnes, 1968), during their foliated period the interception loss from ash and beech is still less than that from conifers: values of p and r_a for coniferous forest would both be smaller. The percentage interception loss figures for ash and beech in Table 2 for the foliated period are also less than the ~35% interception loss typical of conifers (Figure 1).

To extrapolate the interception loss measurements at these sites to other locations a simple empirical model of daily interception loss (requiring only daily precipitation data) with two estimated parameters can be used. Provided the model is applied to situations in which similar species are growing in similar rainfall climates it should provide reasonably accurate estimates of seasonal interception loss. The model is represented by

$$I_D = \gamma (1 - e^{-\kappa R}) \qquad (6)$$

where I_D, the daily interception loss (mm), is a function of R the daily rainfall (mm) and γ (mm) and κ (mm^{-1}) are the optimised interception parameters. Equation (6) was used with 30 months of daily rainfall values for Black Wood assuming that the trees were without leaves between 20 October and 1 June each year.

On an annual basis the percentage interception loss from the beech estimated using (6) for 1989/90 and 1990/91 at 13.6% and 16.2% is in agreement with the values determined from the Black Wood throughfall and stemflow gauges of 13.7% and higher than the 7.3% measured interception loss from the ash (Table 2 and Figure 1).

SIMPLE DAILY WATER-USE MODEL

The more physically realistic water-use models generally require detailed information on surface cover and climate, typically meteorological data measured above the forest canopy on an hourly or more frequent basis, and surface conductances, leaf area and canopy cover etc. Such data are rarely available, except at research sites, and models requiring fewer data are of greater utility for operational purposes. There is however inevitably more empiricism in these simple models which limits the range of their applicability.

The simple water-use model works on a daily time step with an input of daily rainfall and Penman E_T. The model was tested against soil moisture measurements at Black Wood and Old

Pond Close. The procedure follows that of Calder et al. (1983) and Harding et al. (1992a). The soil moisture deficit on day i+1 (δs_{i+1}) is calculated from the soil moisture deficit on day i (ds_i) and the rainfall and evaporation on day i (R_i and E_i)

$$\delta s_{i+1} = \delta s_i + E_i - R_i$$

if δs becomes negative it is put to zero after one day. E_i is calculated as the sum of the daily interception and transpiration from deciduous forest described above. The predicted values of δs were compared with the measured values

$$\delta s_n = s_0 - s$$

where s is the measured total water content and s_0 is the field capacity. For the chalk soils at Black Wood it is necessary to include an additional drainage function which allows drainage from the base of the soil profile (D) even when there is a soil moisture deficit:

$$D = A e^{-B\delta s}$$

where A and B are empirical parameters optimised for each plot.

The equations relating the daily interception loss to the daily rainfall (6), and the daily transpiration to the daily E_T value (2) to (4), were used with the appropriate weather station data. The soil-moisture model parameters were optimised using the sum of squares of the difference between the model prediction and the measurement of soil moisture for the individual tubes as the objective function.

LONG-TERM WATER USE CALCULATIONS

To assess long term effects of broadleaved plantation the model was run for a 20 year period using standard climatological records. A 20 year record (1967 - 1986) of daily Penman E_T values were taken from a climatological station in the Grendon Underwood experimental catchment, Buckinghamshire [51° 50¢ N, 0° 55¢ W]. Because Penman E_T does not vary greatly over lowland England, this record was used for calculations for both Northhamptonshire and Hampshire. The record encompasses a wide range of conditions including the severe drought of 1975/76. The long term calculations were made for the Black Wood area using a daily rainfall record from Overton, 8 km to the north of Black Wood. For the Northampton area the rainfall record from Yardley Hastings was used (the rainfall records starting in 1970). Other variants of the model (Harding et al., 1992) were used to estimate the water use of winter wheat and conifers for comparison; grass evaporation was assumed to equal E_T.

Table 3 The mean annual rainfall and mean estimated annual evaporation for 1967-1986

	Rainfall	Grass	Ash	Beech
Chalk (Hampshire)	740	468	363	423
Clay (Northamptonshire)	589	418	374	-

Mean annual totals of the rainfall and total evaporation for both locations are given in Table 3 for 1967 to 1986. The mean annual E_T over this period was about 80 mm lower than over the 1989 to 1991 period and the actual evaporation estimates are correspondingly reduced.

The evaporation estimates for the ash and the beech are less than from the grass. The comparatively low tree evaporation is a consequence of the strong stomatal control on transpiration, the low interception and the low, or absent, transpiration in the winter. The comparatively high grassland evaporation is an effect of high available water in the chalk soil which leads to a high root constant (K = 160 mm) and results in the grass evaporating at the potential rate in most years.

Water quality

The results of measurements of the solutes in the rainfall, throughfall and stemflow for the beech plantation at Black Wood showed that the trees capture droplets and particulate matter from the atmosphere very efficiently, especially at the forest edge, thereby enriching the composition of stemflow and throughfall. Concentrations and fluxes of most solutes were increased by between 50 and 100% (for example the concentration of nitrogen species (NO_3-N and NH_3-N) was increased from 1.0 to 2.3 mg l^{-1} and that of chloride from 7 to 12 mg l^{-1}). Of the elements measured only strontium, barium and H^+ were not enhanced.

The chemistry of groundwater at Black Wood was determined by extracting water from 21 cored boreholes at depths ranging from 5 m to 30 m in the Chalk unsaturated zone.

There were wide ranges in solute concentrations which were most notable for NO_3, Cl, Na, and SO_4. These solutes are particularly sensitive to environmental influences such as the interception of solutes from the atmosphere, plant uptake and the amount of evaporation. Therefore they are the solutes expected to be most affected by changes in vegetation and land use and least affected by water-rock interactions.

NO_3-N concentrations in the pore water varied from less than 0.1 mg l^{-1} to more than 35 mg l^{-1}. This can be compared with NO_3-N concentrations leaching from undisturbed grassland which are normally in the range 1-2 mg l^{-1} NO_3-N. Excluding samples from the top two metres of chalk which reflect seasonal influences, the average concentration of NO_3-N beneath beech trees in the interior of Black Wood was 9.3 mg l^{-1} compared with 1.2 mg l^{-1} beneath ash trees. These are equivalent to losses of approximately 25 and 3 kg N ha^{-1} a^{-1}, respectively. The difference between the beech and the ash may partly reflect the presence of a dense understorey beneath the ash. Close to the exposed westerly edge of Black Wood, NO_3-N concentrations beneath beech averaged only 1.4 mg l^{-1} but on the less exposed easterly edge NO_3-N concentrations varied considerably with depth and averaged 13 mg l^{-1} NO_3-N.

Nitrate concentrations of between 15 and 40 mg l^{-1} NO_3-N were found in the groundwater beneath clearings within the beech where trees had been felled or had suffered windthrow. These open areas were often colonized by grass and nettles in contrast to the bare areas beneath solid beech stands. The high nitrate concentrations reflect the breakdown of the relatively efficient cycling of nitrogen by the trees and the release of some of the nitrogen that has accumulated in the soil during the growth of the trees. This release appears to continue for more than a decade. The average NO_3-N concentration in these open areas was 21 mg l^{-1} which is of the same order as the concentrations in groundwater draining fertilized agricultural land. Losses from these open areas are of the order of 60 kg N ha^{-1} a^{-1}.

There was a large increase in the concentration of some solutes in the pore water close to the edge of Black Wood. Chloride concentrations were found of nearly 1000 mg l^{-1} and sulphate concentrations of 400 mg l^{-1} . These concentrations compare with 20 and 40 mg l^{-1}, respectively, typical of the centre of the wood. The concentrations were greatest at the extreme western edge and were not significantly enhanced at distances greater than 30-50 m in from the edge. These unusually high concentrations are thought to reflect the additional inputs of sea salts, pollutant gases and aerosols at the edge and confirm the enhanced inputs at the edge observed in the throughfall chemistry. Part of this increase in solute concentrations in the groundwater is probably also due to greater evaporation at the forest edge but this contribution has been difficult to quantify precisely.

Discussion and conclusions

In recent years plant physiological studies have indicated that ash is able to tolerate a wide range of soil water conditions. Besnard and Carlier (1990) measured maximal values of g_s of 640 mmol m^{-2} s^{-1} for ash growing on the flood plain of the Rhône in France where the water table remained close to 1.9 m depth. Under such conditions it would be expected that the annual transpiration would exceed the values estimated for Black Wood. These high values of g_s contrast markedly with the measurements of Carlier et al. (1992) on ash growing in the French Alps where they were subject to water stress. Under "good water supply conditions" the maximal g_s reached 200 mmol m^{-2} s^{-1}. Roberts and Rosier (1994) report maximal g_s of about 300 mmol m^{-2} s^{-1}. Their measurements at Black Wood showed that on chalk-based soils the stomatal conductances of the ash and the beech did not decrease over the summer despite large soil moisture deficits (~350 mm) developing. In contrast the measurements on the ash at Old Pond Close did indicate a reduction in stomatal conductance, and hence transpiration rate, through the summer with the development of a soil moisture deficit. These two sites are representative of two major geologies in the UK and it is likely that the transpiration estimates reported by Harding et al. (1992) and Roberts and Rosier (1994) are representative of much broadleaved woodland in the UK. For such woodland transpiration is conservative and reasonably predictable using simple models of the type described above across a wide area. However, this generalization is not universal. Poplar and willow which currently are not planted extensively but which are likely to be planted increasingly, particularly in the context of energy forestry, are believed to be profligate water consumers. Swedish work has shown the ability of willow to use large amounts of water as irrigated energy coppice and initial measurements on coppiced poplar in England tend to support these. They are also consistent with measurements by several workers (e.g. Ceulemans et al., 1978; Hansen, 1988; Braatne et al., 1992) of high stomatal conductances in poplar and physiological responses consistent with high water use.

Variation of interception loss between species and age classes make accurate prediction of the interception loss a problem. One empirical approach, but one that would require significant resources, would be a measurement program to correlate the tree age of major plantation species with fractional canopy cover, leaf area, leaf dimensions and tree height. If measurements were also made of canopy storage capacity as a function of leaf area then it should be possible to derive formulae relating tree age to canopy capacity and aerodynamic resistance and fractional canopy cover. These parameters could then be used with, for example, the Gash (1979) interception model with daily rainfall to provide estimates of interception loss.

Acknowledgements

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MODELLING CATCHMENT-SCALE WATER BALANCE DYNAMICS USING LONG TIME SERIES OF RAINFALL, STREAMFLOW AND AIR TEMPERATURE

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Abstract

An outline description is given of a rainfall - runoff modelling methodology having a range of applications for catchment hydrology, including systematic exploitation of information in existing regional databases of daily rainfall, streamflow and air temperature. The technique is demonstrated for a large (894 km^2) catchment monitored as part of the UK national hydrometric network, using data from the corresponding National River Flow Archive and point-rainfall database. Specific applications of the methodology mentioned are (a) transferring information to ungauged catchments using empirical relationships between model parameters and physical catchment descriptors and (b) assessment of the hydrological impacts associated with climate change scenarios.

Introduction

One important component of civil infrastructure is a hydrometric network for measuring and monitoring the movement of water in the environment, e.g. as precipitation, streamflow and groundwater. Information from hydrometric networks assists with water resource system operations and environmental management. In the United Kingdom currently, data from some 1200 river gauging stations and 4900 raingauges provide a good picture of the temporal and spatial variation of rainfall and runoff.

The shorter the water balance period, or the more mechanistic detail in which it is required, the more data-intensive it tends to become. Small catchments (e.g. less than 10 km^2) instrumented for intensive studies may monitor, on an hourly basis, or even more frequently, a wide range of variables, e.g. precipitation, streamflow, soil water content, interception losses and transpiration, groundwater levels and hydrometorological variables for estimating evaporation. While an objective of studies in small catchments is often to improve understanding of the mechanisms and dynamics of hydrological processes, there is an increasing desire to transfer catchment-scale water balance information gained from such studies to larger, less well instrumented, catchments. An attractive proposition is the scaling-up from catchments to regions (cells) with a view towards improving the hydrological component of global circulation models (GCMs).

Techniques are required for transferring (regionalising) information about water balance dynamics between catchments spanning a wide range of physical characteristics. This paper outlines an approach to one important area of the regionalisation problem. A rainfall - runoff modelling methodology (Jakeman et al., 1990; Littlewood and Jakeman, 1994) is demonstrated which, although it employs a very simple model structure and has just six parameters, can give good results for a wide range of catchment size and data time step, e.g. less than 1 km^2 to nearly 10000 km^2 and less than hourly to monthly. The methodology, based on a Unit Hydrograph (UH) representation of the catchment-scale rainfall - runoff process, has been found to perform well for many of the catchments monitored by the UK national hydrometric network (e.g. Littlewood and Jakeman, 1993). It has potential, therefore, to assist with regionalisation of water balance dynamics via empirical relationships between model parameters and physical catchment characteristics (Jakeman et al., 1992). A strong feature of the method is that it identifies a UH for *total* streamflow, rather than only for a 'direct' flow component as in more traditional UH methods. And in many cases the UH can be resolved to obtain its quick and slow flow components. Hydrograph separation and a Slow Flow Index (SFI) analagous to the well known Base Flow Index (BFI) are, therefore, by-products of the method (Littlewood and Jakeman, 1991; 1993). Apart from time series of basin rainfall, streamflow and air temperature (the latter as a crude surrogate for evapotranspiration effects) and catchment area, the methodology requires no other data. Unlike some other (spatially distributed) modelling approaches it does not require information on topography, soils and the channel network.

The focus of interest in this paper is on modelling the *daily* rainfall - runoff dynamics of catchments which have essentially 'natural' flow regimes and are not greatly affected by snow or by groundwater imports or exports (other than by streamflow). Readers are requested to consult Jakeman et al. (1990) and Littlewood and Jakeman (1994) (and references cited therein) for accounts of the methodology (known as IHACRES), its historical developments and the parameter identification algorithm it uses. Those references also discuss the applications of IHACRES for catchment hydrology generally; in this short paper it is possible only to give an abbreviated description of the methodology and its applications.

An outline of the simple model structure and brief reference to the parameter identification method are given in the next section, followed by a demonstration analysis for the Teifi at Glan Teifi, an 894 km^2 catchment which extends from the Welsh uplands towards Cardigan Bay. The paper is concluded with brief comments on the potential of IHACRES for assistance with regionalisation of water balance dynamic behaviour and assessment of the hydrological impacts associated with climate change scenarios.

Model structure

A non-linear conversion of rainfall to 'rainfall excess' (that part of rainfall which leaves the catchment eventually as streamflow) is followed by a linear transformation of rainfall excess to streamflow. The rainfall excess - streamflow part will be described first. Measurement units of rain and flow have been assumed to be the same (e.g. millimetres), unless stated otherwise.

RAINFALL EXCESS - STREAMFLOW

Consider the discrete-time hydrograph such that unit rainfall excess over one data time step produces a streamflow B (< 1) over the same time step (rainfall excess and flow have been zero in all preceding time steps and rainfall excess is zero in all subsequent time steps). In each subsequent time step streamflow is a fixed proportion ($A < 1$) of what it was in the previous time step and thus the flow decays exponentially (at a rate determined by A). The resultant UH is shown in Figure 1 and, because the area under the UH ($B + AB + A^2B + A^3B + ...$) is one unit, eqn. (2) follows from eqn. (1). The shape of the UH is completely defined, therefore, by one parameter (either A or B).

$$B(1 + A + A^2 + A^3 + ...) = \frac{B}{1 - A} \tag{1}$$

$$B = 1 - A \tag{2}$$

Convolution of the UH with rainfall excess (..., u_{k-2}, u_{k-1}, u_k, u_{k+1}, u_{k+2}, ...) to give an estimate of streamflow (..., x_{k-2}, x_{k-1}, x_k, x_{k+1}, x_{k+2}, ...) can be executed by recursive application of eqn. (3). Although $B = 1 - A$, it is helpful (as will be seen) to retain both parameters explicitly as in (3).

$$x_k = Ax_{k-1} + Bu_k \tag{3}$$

In principle, any number of such UH elements in series and/or parallel can be tested using the IHACRES methodology to determine which configuration is best for any particular case. In

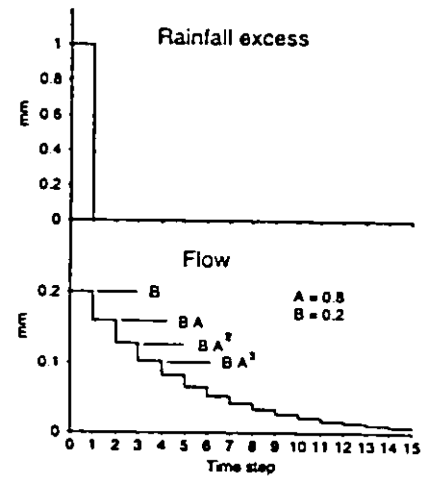

Figure 1 Unit rainfall and Unit Hydrograph

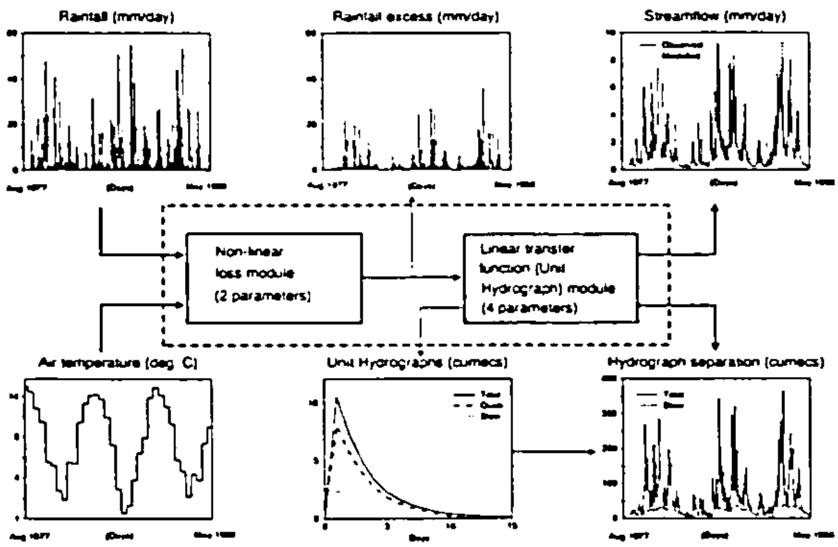

Figure 2 Schematic of the IHACRES modelling procedure

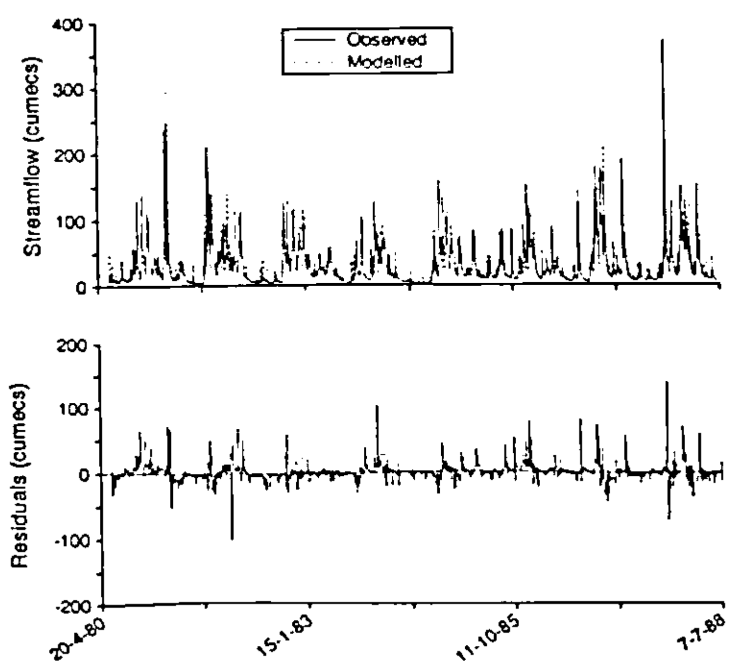

Figure 3 Calibration model-fit 4 June 1980 to 27 June 1988

practice, subject to adequate data quality and a suitable data time step, two UHs in parallel have been found to be optimal for a wide range of catchments. The rates of exponential decay of the two UHs identified are relatively 'quick' (q) and 'slow' (s). In this case, eqns. (4) and (5) apply.

$$\frac{B^{(q)}}{1 - A^{(q)}} + \frac{B^{(s)}}{1 - A^{(s)}} = 1 \tag{4}$$

$$x_k = A^{(q)} x_{k-1}^{(q)} + A^{(s)} x_{k-1}^{(s)} + B^{(q)} u_k + B^{(s)} u_k \tag{5}$$

For purposes of illustration, assume that the areas under the quick and slow UHs are 0.4 and 0.6 respectively, and that the decay rates of the quick and slow UHs are determined by values for $A^{(q)}$ and $A^{(s)}$ of 0.6 and 0.8 respectively. It follows from eqn. (4) that $B^{(q)}$ and $B^{(s)}$ are 0.16 and 0.12 respectively. Note that the area under the slow flow UH ($B^{(s)}/(1 - A^{(s)})$=0.6 in this case) is a Slow Flow Index SFI comparable to the well known Base Flow Index BFI (e.g. Gustard et al., 1992). Characteristic time constants for the quick and slow UH decays can be calculated from eqns. (6) and (7), where Δ is the data time step.

$$\tau^{(q)} = \frac{-\Delta}{\log_e A^{(q)}} \tag{6}$$

$$\tau^{(s)} = \frac{-\Delta}{\log_e A^{(s)}} \tag{7}$$

This 'two UHs in parallel' model structure accords well with the observation that streamflow hydrographs are the superposition of long (slow) recessions during periods of no rainfall excess and (quick) runoff responses caused by rainfall excess events. Indeed, as will be seen later, convolution of a slow flow component UH with rainfall excess can give a reasonable 'baseflow' hydrograph separation.

RAINFALL - RAINFALL EXCESS

The rainfall - rainfall excess part of the model (the 'loss' module) accounts for the non-linearity in the catchment-scale rainfall - runoff process. Its key features in this paper are that it has just three parameters (t_w, C and f) and that it uses only rainfall and temperature data (r_k and T_k (°C)).

The observation that usually, in response to the same amount of rainfall, an antecedently wet catchment produces more streamflow than a dry one, is employed in the form of a catchment wetness index s_k $(0 < s_k < 1)$; rainfall excess u_k is calculated as the product of r_k and s_k (or the average of s_{k-1} and s_k). Several variants of a basic loss module have been tried; eqn. (8a-c) works well on a range of catchments.

$$u_k = \frac{r_k(s_{k-1}+s_k)}{2} \tag{8a}$$

$$s_k = Cr_k + \left[1 - \frac{1}{\tau_w(t_k)}\right] s_{k-1} \tag{8b}$$

$$\tau_w(t_k) = \tau_w \exp(f(T - t_k)) \tag{8c}$$

Parameter τ_w in eqn. (8c) is the value of $\tau_w(t_k)$ at a reference temperature T. In eqn. (8b), $\tau_w(t_k)$ controls the rate at which s_k decays in the absence of rainfall. For time intervals *with* rain, this decay still occurs but s_k is also incremented by a proportion (C) of r_k. A value for C in eqn. (8b) is selected (i.e. calculated automatically during model calibration by computer program) such that the volumes of rainfall excess and observed streamflow over the model calibration period are equal. Model calibration periods are selected, therefore, to start and finish at times of low flow (e.g. at suitable times near the end of the regional water-year) such that the net change in catchment storage of water over that period is close to zero. Parameter f in eqn. (8c) controls the sensitivity of $\tau_w(t_k)$ to changes in temperature.

The whole model, therefore, commonly has six parameters; any three of $A^{(q)}$, $A^{(s)}$, $B^{(q)}$ and $B^{(s)}$, plus f, τ_w and C. A pure time delay **d** (an integer multiple of the data time step) can be applied, in which case the u_k's in eqns. (3) and (5) become u_{k-d}'s. Figure 2 is a schematic of the modelling procedure (note that, as indicated in Fig. 2, good results can be obtained using monthly temperatures with daily rainfall and streamflow). The choice of loss module in this paper (8a-c) is not definitive (see Concluding Remarks).

Parameter identification

A key to the success of IHACRES is its method of identifying the parameters in the linear part of the model (the A's and B's). The Simple Refined Instrumental Variables (SRIV) method in IHACRES can identify the A and B parameters of a rainfall excess - streamflow model structure of the type described above where other selected parameter optimisation algorithms either fail or return sub-optimal results (Jakeman et al., 1990).

Optimal values of f and $\mathbf{t_w}$ can be considered to be those which give a best trade-off between a high coefficient of determination (D) for observed and modelled flow and a low 'average relative parameter error' (ARPE) for the A and B parameters (details are given by Jakeman et al., 1990 and Littlewood and Jakeman, 1994). The search for this trade-off point is facilitated by repeatedly running a model calibration computer program with different values of f and $\mathbf{t_w}$.

TEIFI AT GLAN TEIFI

Daily streamflows for the Teifi at Glan Teifi were obtained from the National River Flow Archive at Wallingford and daily catchment rainfalls were derived from the records of raingauges in and around the catchment using a 'triangular method' of spatial averaging (Jones, 1983). Monthly mean air temperatures were obtained as 40 km by 40 km MORECS cell values (Meteorological Office, 1982). Sixteen overlapping periods, each of approximately 3 years, were selected for analysis. 'Best' models were obtained for each period based mainly on a trade-off between D and ARPE but with attention also to achieving low values for bias and for the cross correlation coefficients between r_k and x_k (x1) and between u_k and x_k (u_1). Some subjectivity appears to be inevitable in such an exercise. The results (Table 1) indicate that better models can be derived for some periods than for others.

Although ARPE was low for all periods (0.03 to 0.07), Model 12 gave the highest D (0.884) and the lowest bias (0.05) while Model 3 gave the lowest D (0.782) and the highest bias (2.73). Note the association between relatively high biases and relatively high cross correlation coefficients. However, a model with low calibration bias can perform well when applied in simulation mode to a period for which the calibrated model has a high bias. For example, Models 12 and 13 (i.e. those calibrated on periods 12 and 13 - biases 0.05 and 0.08 respectively) gave biases of 0.27 and -0.07 respectively when applied in simulation mode to period 1, compared to a calibration bias of 1.83 for Model 1. The simulation mode values of D for Models 12 and 13 on period 1 (0.758 and 0.764 respectively) were not as high as the calibration mode value of D for Model 1 (0.824). This illustrates the need to look at several diagnostic statistics for several sub-periods of the available record when searching for a model which best characterises the catchment.

Table 1. Calibration model-fit statistics

Period	Date	f	t_w (T=10°C)	% Runoff	D	Bias	x1	u1	%ARPE	C
1	15/9/71-1/6/74	0.0248	18	75	.824	1.83	2184.3	-263.7	0.04	0.0125
2	19/10/72-15/8/75	0.0744	38	74	.820	1.60	-39.9	3.0	0.04	0.0131
3	4/9/73-19/8/76	0.0744	14	73	.782	2.73	617.7	-72.6	0.07	0.0306
4	1/6/74-14/8/77	0.0744	34	68	.805	0.75	92.7	-7.2	0.07	0.0152
5	15/8/75-20/6/78	0.1178	26	69	.858	0.55	67.5	-8.7	0.06	0.0303
6	19/8/76-25/7/79	0.1054	26	72	.863	0.17	-90.6	14.5	0.05	0.0254
7	14/8/77-4/6/80	0.0744	26	76	.861	0.73	155.1	-22.5	0.04	0.0180
8	20/6/78-13/8/81	0.0744	34	74	.834	1.61	222.0	-28.5	0.04	0.0145
9	25/7/79-8/8/82	0.0744	26	79	.834	0.88	-25.6	3.4	0.05	0.0173
10	4/6/80-3/8/83	0.0744	50	72	.855	0.35	58.3	-7.1	0.04	0.0101
11	13/8/81-17/8/84	0.0744	46	71	.872	-0.45	663.8	-78.2	0.04	0.0106
12	8/8/82-3/6/85	0.0992	38	69	.884	0.05	-148.3	19.8	0.05	0.0177
13	3/8/83-26/9/86	0.0806	26	67	.847	0.08	47.4	-6.7	0.05	0.0191
14	17/8/84-22/8/87	0.0744	34	71	.861	0.38	572.8	-76.8	0.03	0.0138
15	3/6/85-27/6/88	0.0744	38	73	.858	1.57	-6.6	2.6	0.03	0.0124
16	26/9/86-22/7/89	0.0744	30	78	.830	1.32	796.3	-156.8	0.04	0.0159

Provided a characteristic of the underlying dynamic rainfall - runoff behaviour of a catchment (e.g its Unit Hydrograph) remains stable through time it can be expected that better model parameters will be obtained using longer, rather than shorter, model calibration periods. A 'best' model identified for the sub-record spanning overlapping periods 10 to 15 (about 8 years) gave values for D, %ARPE and bias of 0.855, 0.01 and 0.34 respectively compared, for example, to 0.884, 0.05 and 0.05 respectively for Model 12 (about 3 years). The reduction in %ARPE from 0.05 to 0.01 indicates the improvement in precision on the A and B parameters obtained by using the longer period of record. While D and bias for the model calibrated on the longer period are inferior to those for Model 1 it appears to be a sacrifice worth making. Figure 3 shows the model-fit over the longer calibration period.

The non-linear part of the model simulates rainfall excess from rainfall and temperature using catchment wetness index s_k which should vary between zero and 1. Figure 4a shows, however, that in winter periods s_k actually rises above 1, leading to rainfall excess u_k being greater than rainfall on some occasions (Fig 4b). Occasional $u_k > r_k$ are tolerable on the basis, for example,

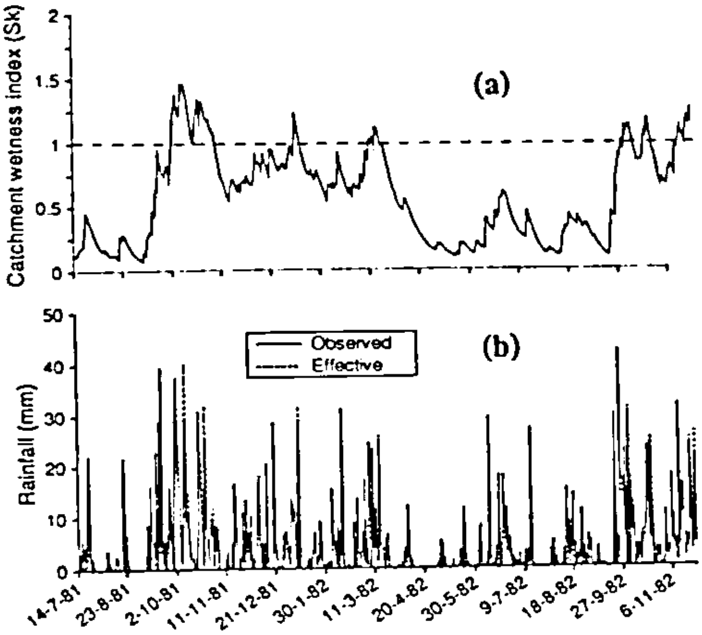

Figure 4 July 1981 to November 1982 (a) Catchment wetness index s_k, (b) Observed and excess rainfall

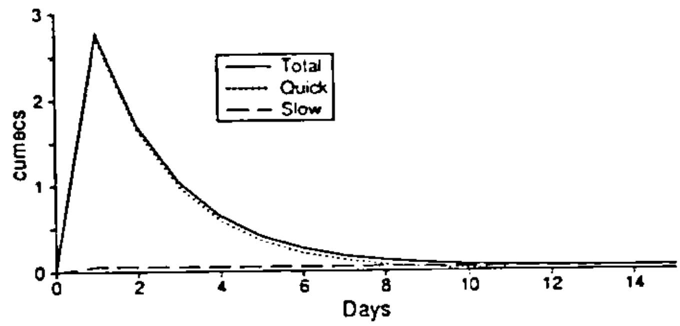

Figure 5 Unit Hydrographs for total, quick and slow flow for model calibrated 4 June 1980 to 27 June 1988

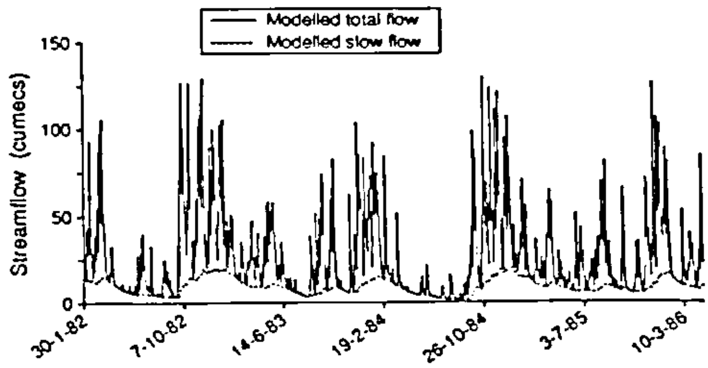

Figure 6 Hydrograph separation January 1982 to March 1986

that estimated basin rainfall may sometimes be less than actual basin rainfall but sequences of $u_k > r_k$ would be less acceptable.

The UH for total streamflow corresponding to the model derived for the approximately 8 year period, and its quick and slow flow UH components, are shown in Fig 5. The hydrograph separation effected by convoluting the total and slow UHs separately with rainfall excess is shown in Fig 6 for about 4 years.

Concluding remarks

One major area of interest currently is regionalisation of simple water balance models to investigate the hydrological impacts which would occur if the climate were to change (e.g. Arnell and Reynard, 1993). The potential of IHACRES in the context of regionalising streamflow dynamic aspects of a water balance, via empirical relationships between model parameters and physical catchment descriptors, has been discussed by Jakeman et al. (1992) and Littlewood and Jakeman (1994), and investigated preliminarily by Sefton et al. (1993). IHACRES models can also be employed to investigate the impact of modest climate changes (rainfall and temperature) on the mean streamflow of gauged catchments (Jakeman et al., 1993).

The rainfall excess part of IHACRES given by eqn. (8a-c) in this paper is one which has been found to work well for a number of catchments in different hydroclimatological zones but it may not be the best loss module for any given catchment or application. Some work to investigate the performance of different loss modules has been undertaken (Littlewood and Post, 1993; Chen et al., 1993) but further work is required to investigate other rainfall excess modules (e.g. in which the evaporation process features more explicitly) which may be more suitable for specific studies.

Acknowledgements

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SPATIAL VARIABILITY AND ACCURACY OF REMOTE SENSING ESTIMATES OF EVAPORATION AND SOIL MOISTURE WITHIN EFEDA'91

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Abstract

The land state variables and fluxes that are derived from remote sensing are very important in hydrological and climatological modelling. For two of these variables and fluxes, namely latent heat flux and soil moisture content the method of retrieval and moreover the accuracy of estimation has been validated. This has been done for the Barrax area in Spain using data collected during the EFEDA'91 experiment. Considerable variations in parameter values exist and deviations from single local measurements can become very large. However, comparison with a combination of various local measurements over a limited number of pixels show an improvement.

1. Introduction

Owing to the spatial heterogeneity of soil physical properties, vegetation cover and precipitation (incl. irrigation), the soil moisture status and evaporation vary spatially. Significant variability of latent heat fluxes is observed at length scales from fractions of a meter to hundreds of kilometers (Feddes et al., 1993). The non-linearity between latent heat flux and soil moisture, and between soil moisture and soil properties implies that in principle straightforward averaging of soil moisture over land surface elements does not yield the correct integrated areal average of latent heat fluxes. If the actual spatial distributions of the latent heat flux were known, the integrated areal average could be calculated correctly. At present a consensus does not exist on how best to approximate the integrated areal average with insufficient knowledge of the areal distribution of the latent heat flux. The calculation of these averages is necessary to obtain data at increasing length scales, as required for many applications, e.g. regional hydrology and climate modelling.

The use of distributed models of land surface processes offers the possibility of obtaining integrated areal averages at the required length scale. At some stage, however, measurements are needed, e.g. model validation. To describe the natural variation of latent heat flux experimentally a multiple set of micro-scale measurements conducted simultaneously at various points is necessary. Such efforts were realized during the EFEDA'91 measurements campaign (Bolle et al., 1993).

Remote sensing instruments can observe a number of locations simultaneously and therefore provide an attractive method to observe the spatial variability of land surface processes at scales which cannot be covered by conventional measurement techniques. Another advantage of remotely scanning devices is that large regions (10-1000 km) are scanned with 100% coverage and different horizontal resolutions can be selected. Menenti (1993) has recently provided an overview of techniques to assess latent heat fluxes from remote sensing measurements in a synergetic way, taking advantage of the interdepency of detectable land surface features.

This paper describes the use of retrieval algorithms for latent heat flux and soil moisture to obtain estimates of the spatial variability of latent heat fluxes and soil moisture under cloud free conditions, within EFEDA'91 and presents an error budget of these method

2. Methods

2.1 Estimation of latent heat flux and of soil water content with remote sensing.

ESTIMATION OF SURFACE FLUXES WITHIN EFEDA

A new type of remote sensing flux parameterization scheme, i.e. the Surface Energy Balance Algorithm for Land (SEBAL), using surface albedo, vegetation index and surface temperature synergetically, has been developed to calculate instantaneous heat fluxes at the land surface. The procedure is based upon a soil heat flux/net radiation fraction, a one-layer resistance scheme for sensible heat flux and a closure of the surface energy budget. The algorithm takes advantage of the relationships between land surface variables existing in nature. This is done using semi-empirical relationships to reduce the number of independent land surface properties. Spatially variable surface properties, such as surface roughness and thermal infrared emissivity, are estimated in this way using e.g. the Normalized Difference Vegetation Index as a prognostic variable. This procedure requires fewer concurrent field measurements for the calculation of heat fluxes. Additional equations for parameter estimation are obtained using simplified forms of the heat balance equation which apply to specific image elements, e.g. where evaporation is either zero or at the full potential rate. Latent heat flux, λE, can in this way be estimated for a wide range of spatial (10m - 5km) and temporal (30 minutes - 16 days) resolutions ranging from small to large regions. For additional details see Bastiaanssen (1994).

ESTIMATION OF SOIL MOISTURE WITHIN EFEDA'91

The microwave region of the electromagnetic spectrum (wavelength between 1 mm and 1 m) has the potential for direct soil moisture estimation by means of remote sensing. Active and passive microwave remote sensing techniques are both used to estimate soil moisture in the upper layer of the soil. The depth of this layer depends upon the wavelength used and the dielectric properties of the soil. In general, the longer the wavelength the larger the depth over which one receives information.

Passive microwave remote sensing, based upon detection of the emitted microwave radiation from the earth's surface, has been shown to give reasonable and consistent results (Schmugge et al., 1992, Van de Griend et al. 1992,). One of the major disadvantages of this technique is the low spatial resolution especially when space borne radiometers are used (e.g. C-band, 150*150 km).

Active microwave remote sensing, based upon detection of reflected or scattered microwave radiation from an object illuminated by an artificial source, provides data with a better spatial resolution. However, soil moisture estimation using these types of instrument has not yet been as successful as in the passive case. Some promising results however, have been published (Dobson and Ulaby, 1986, Lin, 1994). To explore the backscatter behaviour of the land surface, a theoretical model is used, i.e. the Integral Equation Method (IEM) (Fung et al., 1992) and the implications of this for soil moisture estimation in this area are discussed. This discussion is limited to the Barrax supersite of EFEDA'91 and applies to bare soil.

The IEM model describes the backscatter coefficient (σ^0) as a function of the surface roughness, soil dielectric properties (expressed as a complex dielectric constant) and incidence angle. The IEM model combines three other models, namely geometrical optics (**GO**), physical optics (**PO**), being known as the Kirchoff approximations, and the small perturbation model (**SPM**) (Chen and Fung, 1988, Oh et al., 1992). Each model has its own range of validity for the surface roughness parameters (see Table 1). The surface roughness can be expressed as the root mean square (rms) of the height differences relative to the wavelength ($k\sigma$, in which $k=2\pi/\lambda$), the autocorrelation length relative to the wavelength *(kl)* and the rms slope ($m=2\ \sigma/l)^{1/2}$ (Borgeaud and Noll, 1993). The IEM model applies to a much wider range of parameter values including those of the component models.

Table 1: The range of surface roughness for which the backscatter models apply (Borgeaud and Noll, 1993).

Model	Applicable range		
	kl	*kσ*	*m*
Geometrical Optics	> 6	> 2	-
Physical Optics	> 6	-	< 0.25
Small Perturbation	< 3	> 2	< 0.3

2.2 Accuracy of estimates

Given a random variable X, in our case either latent heat λE or moisture content θ, which is a function of the input parameters x_i with $i=1,2,...n$:

$$X = f(x_1,...x_i,...x_n) \tag{1}$$

and a set of reference input parameters of x_i, called r_i, determined from field observations which give as a result the reference value of X hereafter called R:

$$R = f(r_1,...r_i,...r_n) \qquad (2)$$

The input parameters are assumed to be independent and to have a random error that is normally distributed with the reference value as a mean and the standard deviation set at a percentage of the mean. Using sets of the randomly distributed input parameters to simulate errors in input parameters, output for X can be generated from which we can generate the probability density function of X. A large number of realizations of X is neccesary to retrieve meaningfull statistics. One of the important conditions is that the number of realizations must be large enough in order that the statistics become constant. This simulation method is generally known as the Monte-Carlo method.

However, we are not only interested in the distribution of X but also in the properties of the distribution of the differences X-R, i.e. the deviations from the reference value. Thus, the error S_e of X is :

$$S_e = \sqrt{(X - R)^2} = X - R \qquad (3)$$

If we take the absolute value of X-R, we can calculate the cumulative probability density function for the absolute deviations from the reference values, or the error when the reference value R is considered to be the true value for X.

In the case where a spatially distributed random variable X_m is given for j= 1,2,..., m locations then:

$$X_m = f(x_{1,1},...x_{i,j},...x_{n,m}) \qquad (4)$$

and R_m, the reference value of X_m:

$$R_m = f(r_{1,1}, \dots r_{i,j}, \dots, r_{n,m}) \tag{5}$$

In this case the error of estimates S_e of X_m can be calculated with:

$$S_e = \sqrt{\frac{1}{m}\sum_{j=1}^{m}(X_m - R_m)^2} \tag{6}$$

In this paper we assumed that the reference value is the true value for X. However, the reference value is determined using reference input parameters, which were considered to be the true values. This implies that the true value of X is only retrieved when a perfect model is used. In practice better estimates of the term values are obtained using accurate measurements instead of models.

3. Spatial variability

3.1 Latent heat flux

To indicate the extent of spatial variations in λE over the EFEDA'91 grid, the SEBAL algorithm has been implemented at various scales using the image material presented in Table 2. The NS001 is a Daedalus AAD1268 twelve channel multispectral scanner developed at NASA. Fig. 1 shows the frequency distribution of the NS001-based λE-images.

Fig. 1A exhibits several peaks of λE. Pixels with λE>200 W m^{-2} are irrigated fields being approximately 35% of the area. Fig. 1B refers to the Tomelloso area, which appears drier, having more pixels with low λE values. Tomelloso is typically a vineyard region with soil coverages less than 30% and no irrigation. The integrated areal average of latent heat flux from $7*10^5$ pixels and the standard deviation are provided. The areal average surface energy balances in W m^{-2} at 10.21 G.M.T. read as :

(a) Tomelloso : $\langle Q^{*} \rangle(500) = \langle G_0 \rangle(105) + \langle H \rangle(312) + \langle \lambda E \rangle(83)$

(b) Barrax : $\langle Q^{*} \rangle(544) = \langle G_0 \rangle(83) + \langle H \rangle(302) + \langle \lambda E \rangle(158)$

confirming that Barrax is more intensively cropped with a higher net radiation Q^{*} (albedo relatively low), lower soil heat flux G_0 (shaded soils) and higher evaporation (low surface resistance to evaporation).

Table 2: Overview of image material used to study the spatial variability of evaporation within EFEDA

Image	Area	Date of acquisition	Pixel resolution	Total size
METEOSAT	Iberian Peninsula	June, 29, 1991	$3.8*4.1\ km^2$	$10^6\ km^2$
Landsat-TM	EFEDA	June, 12, 1991	$28.5*28.5\ m^2$	$10^4\ km^2$
NS001	Barrax	June, 29, 1991	$18.5*18.5\ m^2$	$300\ km^2$
NS001	Tomelloso	June, 29, 1991	$18.5*18.5\ m^2$	$300\ km^2$

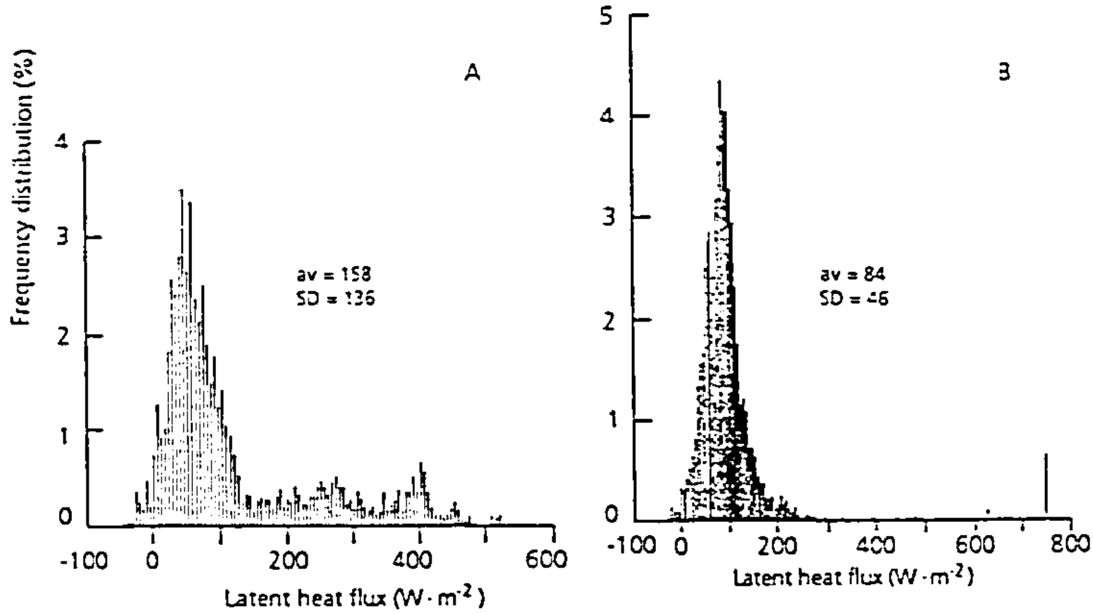


Fig. 1: Spatial variation of latent heat fluxes at the Barrax (Part A) and Tomelloso (Part B) supersites

3.2 Soil water content

For four fields in Barrax the average backscatter for HH and VV polarizations in the L-, and C-band (f = 1 .2 and 5.3 Ghz and λ = 21cm and 5.6cm, respectively) are calculated with the IEM model using the autocorrelation length kl, rms of the height differnces $k\sigma$, the complex dielectric constant (ε'-jε") and the incidence angle (θ_i) as input. The observed (AIRSAR observations) and calculated (IEM model) values of σ^0 for four fields are presented in Table 3. The prediction of the IEM model agrees well with the observations for field 2 (bare soil) and 5 (irrigated maize, in early growing stage) and to a lesser extent for field 4 (full grown alfalfa) and 7 (irrigated barley, senescent stage). However, the IEM model has not been validated well enough yet. This lack of validation, along with the differences in vegetation type and biomass on the fields, limits the use of the model. The trends in backscatter are good - only field 4 with green vegetation shows no correlation with the observations. The model results for the other fields correspond better to the observations, except for field 7 which shows the proper trend, where the values are too low however (see also Bastiaanssen et al., 1994).

Table 3: Results of σ^0 observations and predictions by two models for fields 2/4/5/7 on 19 June, 1991.

	Bands and polarization								
	C-HH	C-VV	C-HV	L-HH	L-VV	L-HV	P-HH	P-VV	P-HV
Observed Field									
2	-13.03	-13.73	-23.34	-21.50	-19.57	-36.68	-25.46	-22.37	-38.34
4	-7.2	-9.52	-15.65	-16.24	-14.55	-28.57	-26.74	-20.59	-35.34
5	-8.64	-9.30	-17.25	-17.05	-15.47	-31.88	-23.46	-21.47	-36.17
7	-12.99	-15.39	-22.86	-24.40	-22.87	-36.47	-28.39	-24.31	-29.50
Predicted									
5	-7.4	-10.0	-	-15.4	-11.9	-	-23.3	-19.3	-
7	-18.4	-19.4	-	-29.6	-24.3	-	-35.8	-29.9	-

4. Accuracy

4.1 Accuracy of estimated latent heat flux

OVERALL ACCURACY

Although promising results for the determination of evaporation through remote sensing observations have been reported (e.g. Schmugge and Becker, 1991), the successful testing of these algorithms at meso-scale is often extremely difficult to realize.The SEBAL-technique was initially applied to map regional evaporation in Egypt, i.e. Qattara Depression (Menenti et al, 1991), Nile Delta (Bastiaanssen et a., 1992). The availability of the EFEDA'91 data base of λE field data provides a good opportunity to investigate the accuracy of the SEBAL algorithm. Values for λE could not be compared directly because of different temporal resolutions of tower-based turbulent heat fluxes and the time scale at which the SEBAL scheme operates. On the other hand the evaporative fraction, Λ, expressed as:

$$\Lambda = \frac{\lambda E}{H + \lambda E} \qquad (7)$$

where H is the sensible heat flux, is a feasible and simple parameter to charcterize the energy budget of heterogeneous land surfaces.

Fig. 2 illustrates the agreement between remotely sensed and tower-based values of these evaporative fractions. The overall accuracy of the remotely sensed Λ is 0.09 whereas the integrated areal average for Barrax and Tomelloso is $<\Lambda>=0.283$. The agreement for the three tower stations on a 1:1 (45°) line looks poor. After applying error bounds on the field measurements however, it was concluded that the SEBAL estimates are in the error range for all individual meteorological stations operated during EFEDA'91. The error range applies to eddy correlation, Bowen-ratio and aerodynamic profile measurement systems.

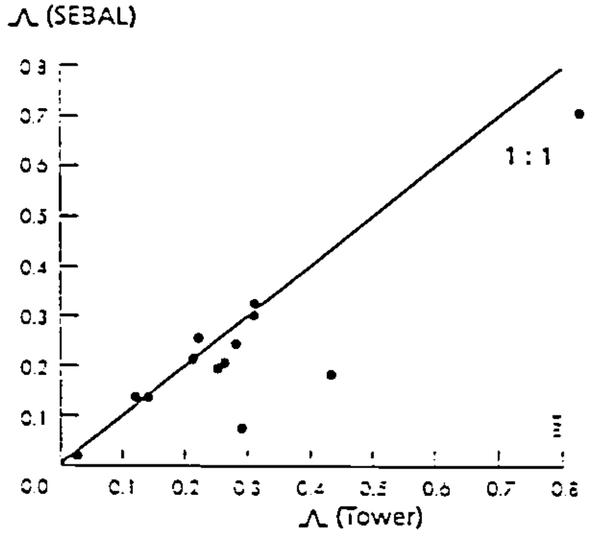


*Fig. 2: Comparison of evaporative fractions, Λ, as estimated with the SEBAL algorithm, using NS001 measurements and in-situ tower measurements considering rectangular fetches of 92.5*92.5 m^2 (5 * 5 pixels), EFEDA'91.*

PROBABILITY DISTRIBUTION FUNCTION OF ERRORS

The lack of knowledge on confidence intervals forced us to study the statistics of the error distributions of the algorithms applied in estimating latent heat flux and soil water content. This approach, as described in section 2.2, is restricted to random errors. The error at homogeneous patches, e.g. a single pixel, may be unacceptably high, while the integrated areal average latent heat flux, consisting of a large number of various homogeneous patches, can be sufficiently accurate. Table 4 summarizes the model parameters $x_i..x_n$ which were considered to have randomly distributed errors in the calculation of the probability distribution function (pdf) for X_m - R_m, where X_m represents the areal average $\langle \lambda E \rangle$-value. The random errors of each model parameter were assumed to have a normal distribution.

Table 4: Parameters of the SEBAL algorithm considered to have a randomly distributed error.

1. Surface albedo	10. Surface roughness for momentum transport
2. Normalized Difference Vegetation Index	11. Surface roughness for heat transport
3. Surface temperature	12. Slope albedo/temperature
4. Incoming longwave radiation	13. Slope longwave radiation/temperature
5. Incoming shortwave radiation	14. Slope soil heat flux/temperature
6. Surface thermal infrared emissivity	15. Windspeed
7. Soil heat flux/net radiation fraction	16. Stability correction for momentum transport
8. Air density	17. Stability correction for heat transport
9. Friction velocity	18. Near-surface vertical air temperature difference

To estimate the error bound on each parameter the standard deviations were set at 5% of the reference parameter value. To account for the non-linearity of the SEBAL algorithm, eq. (6) was applied to 20 different reference locations. Reference values r_{ij} for surface albedo, vegetation index and surface temperature were taken as mean values for each of the 20 sub-units into which the Barrax NS001 image was divided. The reference values for the remaining parameters of Table 4 were taken as the mean values computed with the SEBAL algorithm for the 20 sub-units. Consequently, R_m is based on models, rather than on accurate field measurements. To calculate the pdf for each parameter and location, 5000 pixels * n=18 parameters * m=20 locations = $18*10^5$ random values were generated. The pdf of errors (Fig.3) shows that the error on λE at the 90% confidence level is 38%. If instead of 20 sub-units, 1 sub-unit was chosen, the errors would have been larger (especially at the dry plots).

The selection of error bounds on individual parameters and of the reference values has been somewhat arbitrary and should be improved in subsequent studies. Considering the overall deviation from the 1 : 1 line being Λ=0.09 and $<\Lambda>$=0.283, which means an error in the integrated areal average evaporative fraction of 32%, we conclude that the first estimate of the error ranges in the 18 model parameters of Table 4 are reasonable since 32% error corresponds to 83% confidence level.

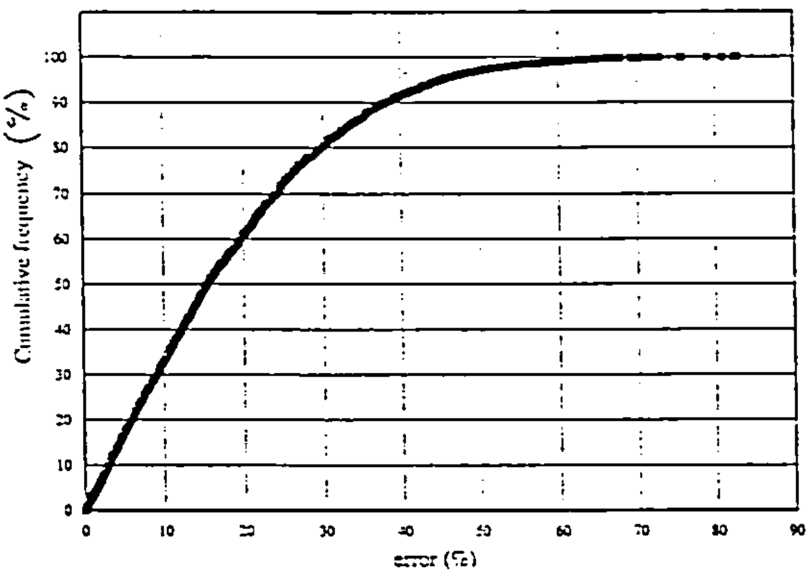


Fig. 3: Probability distribution function of errors for the SEBAL algorithm over the heterogeneous Barrax supersite within EFEDA'91.

4.2 Accuracy of soil moisture estimation

OVERALL ACCURACY

The validation of the soil moisture retrieval algorithms using active micowave remote sensing proves to be very difficult at a number of scales. The most difficult part is to estimate the penetration depth of the radiation, which is the depth over which one estimates the amount of soil moisture but which is also dependent on the amount of soil moisture present. To obtain the total amount of water integrated over the penetration depth is difficult and it requires a large amount of ground data combined with modelling of the soil moisture profile. This makes it tedious to validate the IEM-model and consequently to asses the overall accuracy of soil moisture estimation.

PROBABILITY DISTRIBUTION FUNCTION OF ERRORS

A similar procedure as that used with latent heat fluxes was applied to obtain the pdf of the errors on the estimated backscatter coefficient. The parameters considered as sources of random errors are listed in Table 5.

*Table 5: Parameters of the IEM model considered to have a ran*domly distributed random error.

1.	Autocorrelation length
2.	RMS of height differences
3.	Complex dielectric constant
4.	Incidence angle

The reference values of the model parameters listed in Table 5 are obtained on the basis of field measurements. Next eq.(6) was applied. The error bounds on the individual parameters were estimated by taking the standard deviation as being 20% of the mean for both the autocorrelation length and the complex dielectric constant and 10% of the mean for both the incidence angle and the rms of the height differences.

The pdf of error of estimate (Fig. 4) for VV-polarized C-band indicates that the accuracy on estimated backscatter is about 17% at the 90% confidence level.

To estimate the soil water content the IEM model has to be applied in an inverse manner. This implies that the error in estimating σ^0, combined with a lower sensitivity of σ^0 to soil water content, makes inversion rather challenging.

If instead of the VV-, or HH-polarized backscatter values, the ratio of HH- and VV - polarization is taken, the results improve with narrower distributions indicating less variation. The accuracy in this case is 11% at the 90% confidence level.

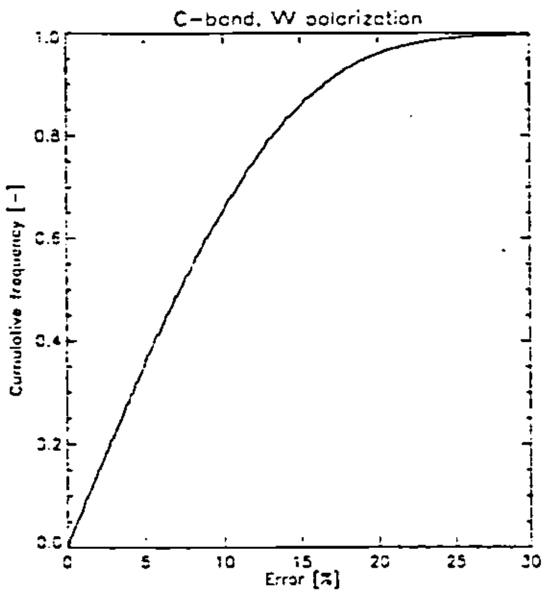


Fig. 4 Overall error-probability function for the backscatter coefficient σ^0 assessed with the IEM model for C-band, VV-polarization.

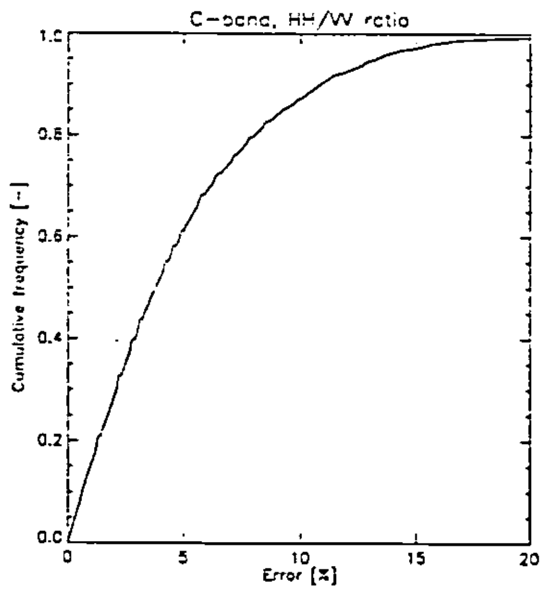
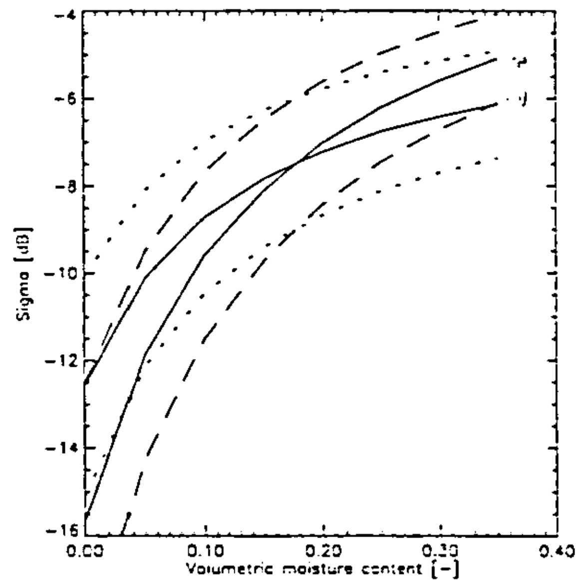


Fig. 5: Overall error-probability function for the backscatter coefficient σ^0 assessed with the IEM model for C-band, HH/VV-polarization.

Fig. 6A illustrates the C-band backscatter coefficient plotted as a function of the volumetric water content (θ) with all the other parameters fixed at the value determined in field 2 (Table 2). As can be seen, there is a non-linear relationship between θ and VV- or HH backscatter. Above about $\theta = 0.20\ cm^3\ cm^{-3}$, this relationship becomes less meaningful. Fig. 6B for the same plot shows the effect of taking the HH/VV ratio, resulting in an almost linear relationship. Similarly both VV- and HH backscatter coefficients are sensitive to surface roughness effects. By taking the ratio of these two, one can eliminate these effects.



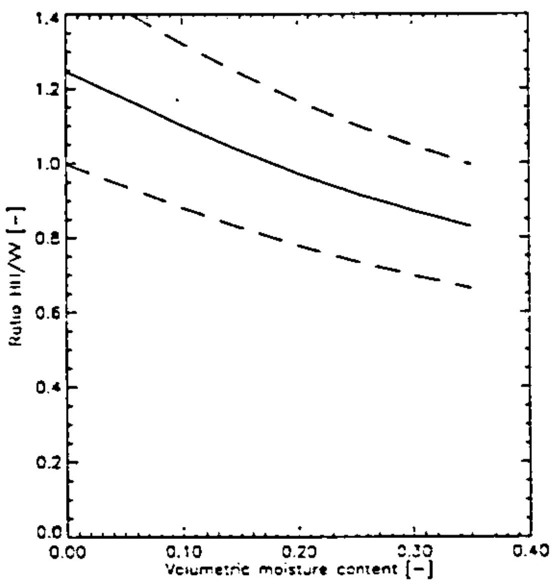


Fig. 6 A) Backscatter coefficient vs. soil water content (HH and VV polarization); B) ratio of backscattter coefficient with VV polarization over HH vs soil water content.

(Solid lines indicate the computed relationship, dotted lines the standard eleviation).

The ratio HH/VV depends almost linearly on volumetric soil water content. In this case, the variation in the backscatter ratio is much less and there is a fairly strong relationship between the backscatter ratio and soil water content. Thus inversion of the model c.q. soil moisture retrieval can yield reasonable results over the whole range. This approach applies only to certain ranges for the roughness parameters and also when the backscatter of the VV or the

HH polarizations is more sensitive to soil moisture than the other. When the geometrical optics approximation is valid (Borgeaud and Noll, 1993) then both polarizations are identical and thus the ratio HH/VV is always one and inherently insensitive to soil moisture increase.

5 Accuracy versus spatial variability

The results of the error analysis provide a basis to assess the quantitative value of the spatial information provided by detailed remote sensing data. In the case of the distributed λE values, an error of 38% of the mean value implies that only part of the observed spatial variability is significant. Likewise, differences in σ^0 between the fields (Table 3) are generally not significant.

The advantage of remote sensing data, especially high resolution data, is that a number of samples (i.e. pixels) can be used to obtain estimates with smaller random errors. This estimation can be performed by averaging over a limited number of pixels (say 10), at the cost of a poorer resolution. This approach can only be applied to homogeneous targets, to avoid mixing measurements which are related to different physical objects.

6. Conclusions

The relevance of the results presented here is limited as the determination of the reference values necessary for the errors study were determined from locally valid field observations and model output rather than directly from field observations. Hence, the range over which the accuracy of the SEBAL and IEM algorithms was explored, is limited. Indeed this range was rather large for both latent heat flux and the soil water content for the composite Barrax site.

At the 90% confidence level, the error in estimating λE from remote sensing (SEBAL) is 38% and the error in estimating the backscatter coefficient is 12%.

The advantage of high resolution data is that a number of pixels can be used to obtain estimates with smaller random errors by degrading resolution. This approach is valid for homogeneous targets only, to avoid the mixing of measurements that relate to different physical objects.

The remote sensing applications described here may have several applications in hydrology and meteorology:

- (re)-initialization of land surface parameterization schemes in meso scale studies;
- validation of distributed parameter hydrological models;
- determination of crop stress with respect to water supply;

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WATER: SOME RELATIONSHIPS BETWEEN QUANTITY AND QUALITY

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Abstract

Weather affects water quality. Heavy rainfall increases *loads* of pollutants reaching rivers and lakes but the increased volume of the receiving waters may result in reduced *concentration* of pollutants. Higher temperatures affect the dissolved oxygen content of water both directly and indirectly. This paper examines some relationships between water quality and quantity in Ireland, with special emphasis on the interaction between weather, agriculture and water quality.

Introduction

Weather affects water quality - both directly and indirectly. The quality of surface waters - rivers and lakes - changes greatly from season to season and may even change during a given day reflecting fluctuations in rainfall or radiation/temperature. Groundwater quality may also be affected but the changes occur more slowly.

Changes in rainfall can have a number of different effects on water quality. Where there is a relatively constant discharge of polluted effluent from a *point source,* as for example from a sewage treatment plant or from an industry which has a licence to discharge to water, increased rainfall increases the volume of flow in the river, thus diluting the *concentration* of pollutants in the receiving water and by definition improves water quality where *concentration* is the standard criterion. This has led to the old adage that "dilution is the solution to pollution". The efficacy of the dilution effect in preventing pollution depends on many factors. In the first instance it depends on the concentration of pollutants in the discharged effluent but it also depends on the volume and velocity of flow of the receiving water. This, in turn, depends on the season as evaporation and evapotranspiration in summer greatly reduce the volume of water in the rivers. In years with severe drought, low flows in the majority of Irish rivers are less than 5% of average flows and in some cases low flows are less than 1% of average flows (McCumiskey, 1991). It is easy to understand that a constant *point source* discharge will have a vastly different effect on water quality depending on whether it reaches the receiving river water during low, average or high flow. The effect of the pollutant also depends greatly on the temperature of the receiving water. High temperatures encourage rapid biological growth, which affect dissolved oxygen concentrations. Besides, oxygen is considerably less soluble in warm water than in cold water.

Prolonged heavy rainfall has several other effects on water quality. It causes storm overflow of urban treatment facilities and over-flow of uncovered farm waste storage facilities thus creating temporary *point services* of pollution, although these are often referred to as *non-point* sources. Heavy rainfall also increases runoff from parking lots and paved areas, increased losses of sediment and nutrients from agricultural and forestry land and erosion of river/stream banks which are all diffuse or *non-point sources* of pollution. Depending on the mix of these occurrences in any one catchment, the complex outcome of a heavy rainstorm may result in decreased, unchanged or increased *concentration* of nutrients in the receiving water. The *loads* of nutrients will at best remain unchanged but are almost always increased. In fact, increased *loads* of nutrients in such circumstances are generally considered as an indication of *non-point source* pollution.

Water Quality in Ireland

The quality of surface water in Ireland is generally very good. This is due to a number of factors. There is relatively high rainfall (mean annual 1150mm) which occurs fairly evenly throughout the year, low density of population (50 persons per km^2) and relatively small numbers of industrial discharges to water. In addition, the largest urban areas including Dublin, Cork, Limerick, Galway and Waterford, which account for about one third of the total 3.5m population, are all located on the coast and do not discharge to inland waters.

There are, however, some disturbing trends in water quality which have been demonstrated in four successive biological surveys of river quality carried out in Ireland since 1971 (McCumiskey and Toner, 1992). The first national survey in 1971 involved 2900km of river channel which set a long-term baseline for quality assessment. The length of channel surveyed has increased greatly since that time to about 13,000 km at the present time but the quality status of the baseline 2900 km showed a number of changes over the four surveys (Fig 1). While serious pollution decreased from 7% to 2% of channel affected,

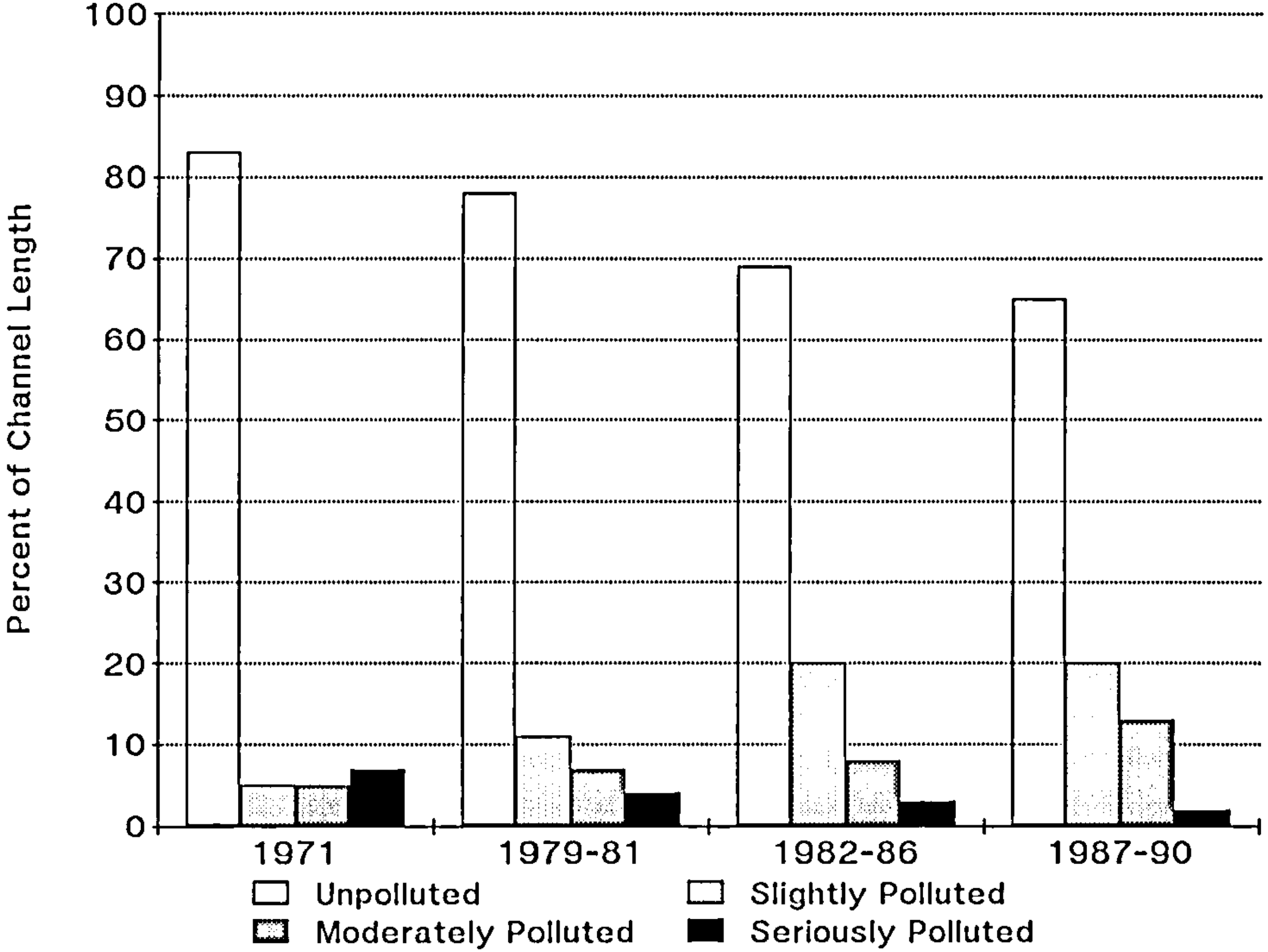

Figure 1. Changes in biological water quality in 2900km of baseline river channel during four successive surveys.

both slight and moderate, pollution increased substantially. Overall, the length of unpolluted water in these stretches decreased from 83% in 1971 to about 65% in the 1987-1990 survey. Slight and moderate pollution are considered to be due mainly to eutrophication arising from inputs of nutrients, particularly phosphorus and nitrogen, from agriculture and to a lesser extent from sewage discharges.

A number of Irish lakes also suffer from eutrophication. Of a total of 164 lakes surveyed (1987 - 1990) 21 had characteristics of eutrophication in varying degrees which was also considered to be mainly due to phosphorus inputs from agricultural activities (McCumiskey and Toner 1992). The reason why phosphorus is singled out as the most polluting nutrient is because it is considered to be the only one of the nutrients essential for plant growth which could be practicably controlled in the ecosystem by limiting inputs. It is generally accepted that proliferation of algae in rivers is limited if phosphorus concentration is less than 30mg molybdate reactive phosphate (MRP) per m^3, but lakewaters should preferably have less than 20mg MRP/m^3 to limit algal growth.

The main purpose in identifying the source of phosphorous *loads* is to quantify the relative inputs from the various sectors so that each sector can examine the possibilities to reduce its contribution.

Agriculture - Point or Non-Point Source Pollution?

The contention that agriculture contributes the largest *loads* of phosphorus and nitrogen to surface waters in Ireland can hardly be disputed. In a recent study of the trophic status of Lough Conn, which has a very rural catchment, McGarrigle et al (1993) estimated that agriculture was responsible for 60% of the phosphorus inputs to the lake during the study period 1990 - 1991. This is not really surprising since in rural areas, agriculture is by far the biggest "industry" and animal populations greatly outnumber human populations. In a similar study of phosphorus inputs to Lough Derg, Bowman et al (1993) also found that more than 50% of the annual phosphate inputs to the lake were comprised of background runoff and agricultural activities.

In order to try to reduce these inputs from agriculture, it is necessary to understand which practices contribute the biggest loads of phosphorus to water and devise guidelines and/or policies to control these practices at farm level.

Sources of Agricultural Phosphorus Reaching Water

The *precise* source of agricultural phosphorus which reaches water is not known with certainty. Most of the possible sources are associated with heavy rainfall and include: runoff from land following landspreading of slurry or phosphorus fertiliser, runoff from soils with elevated levels of soil phosphorus and overflow of farm waste storage facilities. Possible sources which are not necessarily associated with high rainfall are direct discharge of farm wastes, particularly soiled yard water, to ditches or streams, or discharge of milking parlour washings which often contain high concentrations of phosphoric detergents. These latter *point sources* can be remedied by improved farm management practices.

Runoff of Nutrients from Land Following Slurry or Fertiliser Spreading

In a series of experiments stretching over a 6 year period, Sherwood (1990) showed that the most important factor governing the loss of organic matter and nutrients following landspreading of slurry was the time interval which elapsed between spreading and the occurrence of the first rainfall which caused surface runoff. If runoff occurred within the first few days, losses of organic matter (BOD) and nutrients such as phosphorus and nitrogen were all substantial. If 7-10 days elapsed before runoff occurred, BOD concentration of runoff water was greatly reduced, nitrogen concentration was moderately reduced, while phosphorus concentration in runoff water decreased more slowly than either of the other two parameters. Runoff which occurred 3-4 weeks after spreading a normal volume of $35m^3/ha$ of slurry tended to have very low BOD and low soluble nitrogen concentrations in runoff water but phosphate concentration could still be as high as 3000-4000mg MRP/m^3.

These results show the importance of checking the weather forecast before spreading slurry to ensure that heavy rainfall is not imminent and thus avoid losses of large loads of pollutants to surface waters. In Ireland there is a heavy dependence on contractors for slurry spreading which may militate against farmers spreading slurry or fertiliser on dates when weather conditions are optimum. On the other hand, it probably ensures that spreading takes place fairly evenly over the entire season when ground conditions are suitable and in this way spreads the risk of a major catastrophe which could arise if all farmers spread simultaneously throughout the catchment and heavy rainfall or thunderstorms occurred unexpectedly. In the same series of experiments Sherwood (1990) also found that *fertiliser* phosphorus was liable to runoff if heavy rainfall occurred soon after spreading but runoff did not occur over such an extended period compared with slurry.

The foregoing plot results show that there is a *potential* for runoff of phosphorus if spreading occurs when weather conditions are unfavourable, and there is *practical* evidence that this can occur at catchment level and have a detrimental effect on lake water quality. One of the most detailed studies to date on phosphate inputs to an Irish lake in a very rural setting was carried out by McGarrigle et al (1993) who quantified all the phosphate inputs to Lough Conn 1990-1991. This study was particularly interesting as details of inputs to the lake were available from an earlier study (1980-1982) for comparison. The phosphate load reaching the lake, doubled in the 10 year period between studies. A more remarkable change was that there was an obvious response to wet weather events with a positive correlation between volume of water and phosphate concentration in the inflowing waters in 1990-1991, whereas no such relationship was apparent in the earlier study. The "spikes" associated with heavy rainfall were attributed to runoff following landspreading of slurry. In the intervening period, more than 200 slatted floor winter-housing units for cattle had been built, in order to reduce direct discharge of farm wastes to water. If the increased "spikes" were due to farmers landspreading slurry in the late autumn/early winter, in order to have the tanks emptied before housing the animals for the winter, the overall objective of improving environmental protection was certainly not achieved. This occurrence is a good example of the futility of having expensive facilities if there is not a good environmental management system in operation also.

One approach to controlling this type of *non-point source* pollution might be to limit slurry spreading to times/months when runoff is statistically less likely to occur and this is done in some countries. It would be a complicated exercise to devise an equitable system for each county in Ireland as amounts of runoff are very much governed by soil type and slope as well as rainfall. An example of the effect of soil type on runoff volume is shown in data taken from Sherwood and Farming (1981) and presented in Table 1. The volume of runoff water collected from plots on a gley soil was approximately four times greater than the volume collected from a loam soil although the rainfall was similar. The runoff

Table 1: Mean monthly runoff from two soil types, measured on small grassland plots (6 per soil type) 1976 - 1980. Mean rainfall for the period was 1093 mm and mean evapotranspiration was 536mm.

	Runoff Water Collected (mm)	
	Moderately well drained loam	**Impermeable gley soil**
January	16.4	47.5
February	15.7	56.7
March	3.1	25.1
April	0.7	5.1
May	0.3	3.1
June	----	2.6
July	----	----
August	0.4	0.1
September	-----	-----
October	0.3	14.3
November	4.2	27.9
December	9.1	32.4
Totals	50.2	214.8

per month (averaged over 4 years) is also shown. For planning purposes, we might conclude from the data presented, that slurry spreading on gley soils should be prohibited for the months of November - March (inclusive) in areas with about 1100 mm rainfall. However, averages can be misleading. Dry spells can occur in any of these months, in any year, and may present an opportunity to safely spread slurry, particularly in February and March when grass growth has already begun which means nutrient uptake has commenced and slurry nutrients can be used to replace fertiliser nutrients.

Prohibition of slurry spreading in winter months may serve a useful purpose, but it may not be sufficient to prevent pollution. There is a further requirement that farmers should follow codes of good practice during the months when spreading is allowed, taking both weather and soil conditions into consideration.

It should be recognised that most agricultural runoff occurs mainly in winter, when volume of flow in rivers is high and temperatures are low. It may consequently have a much less damaging effect on water quality than *point source* inputs from other sectors which may have their peak discharges in summer. This was highlighted by Bowman et al (1993) in the Lough Derg Report.

Groundwater Quality

The relationship between groundwater quality and rainfall is more complex. Soils and subsoils, depending on their depth and physical composition, act as filters in most cases for infiltrating water. However in many areas, subsoils are thin or absent or there may be karstic limestone features present. In these areas, there is little protection or filtering effect and groundwater is very vulnerable to pollution from septic tanks, leaking farmyard storage tanks and landspreading. Drew and Daly (1993) described many instances of rapid flushing of pollutants to groundwater in karstified regions of Galway, Mayo and North Clare particularly by the first heavy rainfall following a dry period.

Another interesting aspect of the relationship between water quantity and quality is to consider if extra rainfall results in dilution of pollutants in groundwater. In the context of the Nitrates Directive (91/676/EEC) which is intended to control the amount of nitrate reaching water from agricultural sources, it was important to attempt to quantify the amounts of nitrate which might reach groundwater from normal agricultural practices in this country and to assess how these might be influenced by variation in rainfall.

It is generally accepted that leaching of nitrate is higher from tillage crops than from grassland, so it was decided to study the effects of a number of parameters on leaching of nitrate from cereals. For this purpose, lysimeter experiments were carried out by Sherwood and Fanning (1986-1990) at Johnstown Castle, to obtain information on leaching of nitrate from spring barley grown on a loam soil. For 1987 - 1988 the variables included in the study were two nitrogen rates (0 and 120kg N/ha) and three rainfall rates .

Table 2: The effect of different amounts of rainfall on the mean recovery of N in barley and drainage water, volumes and nitrate concentration of drainage water collected from lysimeters which received 0 and 120 kgN/ha in the period April 28, 1987 to May 6, 1988.

	N Recovered (kg/ha)			
Rainfall Treatment	**Plant (Total N)**	**Drainage Water (NO_3^-N)**	**Total Volume Drainage Water (mm)**	**Mean Nitrate (NO_3^-)* Conc. in water (mg/l)**
<u>Barley + 120 kg N/ha</u>				
1144mm rainfall	100	91	785	51
1464mm rainfall	119	114	1090	46
1784mm rainfall	106	109	1452	33
<u>Barley minus Nitrogen</u>				
1144mm rainfall	45	80	832	43
1464mm rainfall	44	111	1174	42
1784mm rainfall	31	121	1500	35
<u>Fallow (no barley)</u>				
1144mm rainfall	0	235	943	110

* The Maximum Admissible concentration for Nitrate (NO_3^-) concentration in drinking water is 50mg/l (80/778/EEC).

This was achieved by having three treatments: (1) ambient rainfall - 1144mm from sowing date 1987 to sowing date 1988, (2) ambient plus 320mm added water and (3) ambient plus 640mm added water. The extra water was added four times weekly from November, 1987 to April, 1988. There was also a treatment with no barley (fallow) which gave a measure of the amount of nitrate leaching from soil nitrogen. Results are presented in Table 2, and show the total amounts of nitrogen (kg N/ha) leached from each treatment, the volume of drainage water collected and the mean nitrate (NO_3^-) concentration in the drainage water for the year between sowing date 1987 and sowing date 1988.

The biggest effect was between barley and no barley. In the absence of a growing plant with no fertiliser added, the amount of nitrate leached was more than double that which leached from fertilised or unfertilised barley. This was a measure of mineralised *soil* nitrogen. There were relatively small differences between the amounts of nitrate leached from fertilised and unfertilised barley at any of the three rainfall rates.

The extra rainfall had a diluting effect on the nitrate in the drainage water but it was not a straightforward *pro rata* dilution. The reason for this was although the nitrate largely came through in a peak with the first 500mm drainage water, the nitrate concentration never dropped to zero - there was always a small amount of background nitrate in the drainage water. However, from a practical viewpoint the results imply that the rainfall difference between east and west in this country could not have a big influence on the nitrate concentration of groundwater under tillage crops.

Conclusions

This paper has attempted to present a few examples of the complex relationship between water quantity and quality in Ireland. These relationships can have practical implications for Local Authorities which have the task of providing water of suitable quality for a variety of beneficial uses. Modern lifestyles pose many conflicting demands. On the one hand, there is increased pressure to provide wealth and jobs and to foster industries and activities, which, depending on their nature, may need licences to discharge polluting effluents to water. On the other hand there are mounting pressures on regulatory authorities to raise standards and increase controls, in order to safeguard drinking water quality and provide water suitable for a variety of amenity uses such as fishing, boating, swimming and even visual impact. It is a considerable credit to the responsible authorities that by and large, they manage to keep all parties reasonably satisfied. However, if there is to be a determined effort made to reverse the trends towards eutrophication, each sector must be educated to understand the significance of the phosphorus loads they discharge, and the influence of meteorological factors in determining whether the pollutant causes serious damage to the ecosystem. Most important of all, all sectors must be encouraged to take ameliorative action.

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HYDROLOGICAL AND LANDUSE CHANGES IN NORTH ANTRIM

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Abstract

The frontal depression of the 27-28 October 1990 was responsible for remarkable rainfall intensities and record floods in northeast Ireland. Intense precipitation had its most dramatic impact in Ballycastle and Cushendall, where flooding was widespread. Extreme precipitation was the causal factor, but the event draws attention to perceived changes of flow regimes and landuse in the upper catchments of rivers in North Antrim. Evidence that human activity in the upland has led to an increased frequency of flooding is found to be inconclusive. A need exists to closely monitor the hydrological characteristics of these catchments in order to calibrate more precisely the flood hazard in this part of northeast Ireland.

Introduction

Precipitation in Ireland is generally perceived as being of low intensity and long duration, with infrequent occurrences of heavy falls due to the relatively low relief, and the limited occurrence of severe convective activity in summer (Betts, 1982). It is salutary to note, however, that 97mm of rain fell in 45 minutes at Orra Beg, North Antrim on the 1 August 1980 - a record for the United Kingdom (Meteorological Office, Belfast, 1988). On the 27-28 October 1990 a vigorous depression crossed Ireland, and intense rainfall over North Antrim culminated in severe flooding and substantial damage, particularly around Ballycastle and in the Glens of Antrim (Figure 1). Extreme precipitation was the causal factor, but the event drew attention to the local community claim of perceived changes in the flow regimes of rivers in North Antrim during the past 2 decades.

Absence of streamflow records and a limited autographic rain gauge network, prevents the quantitative validation of any hydrological change. However, flood alleviation work being undertaken in Ballycastle, and proposed for Cushendall, by the Department of Agriculture, suggests the existence of hydrological problems in North Antrim, although the causes are uncertain. This paper describes the nature of the flood hazard, and considers factors that may be responsible for inducing change in the regime of rivers in the Glens of Antrim.

Study Area

Extending northward from Glenariff, the study area geologically comprises a small inlier of Dalradian schist bounded by Tertiary basalts. Topographically, it forms an upland plateau more than 300m above sea-level, with isolated summits over 500m, and is deeply dissected by glens,

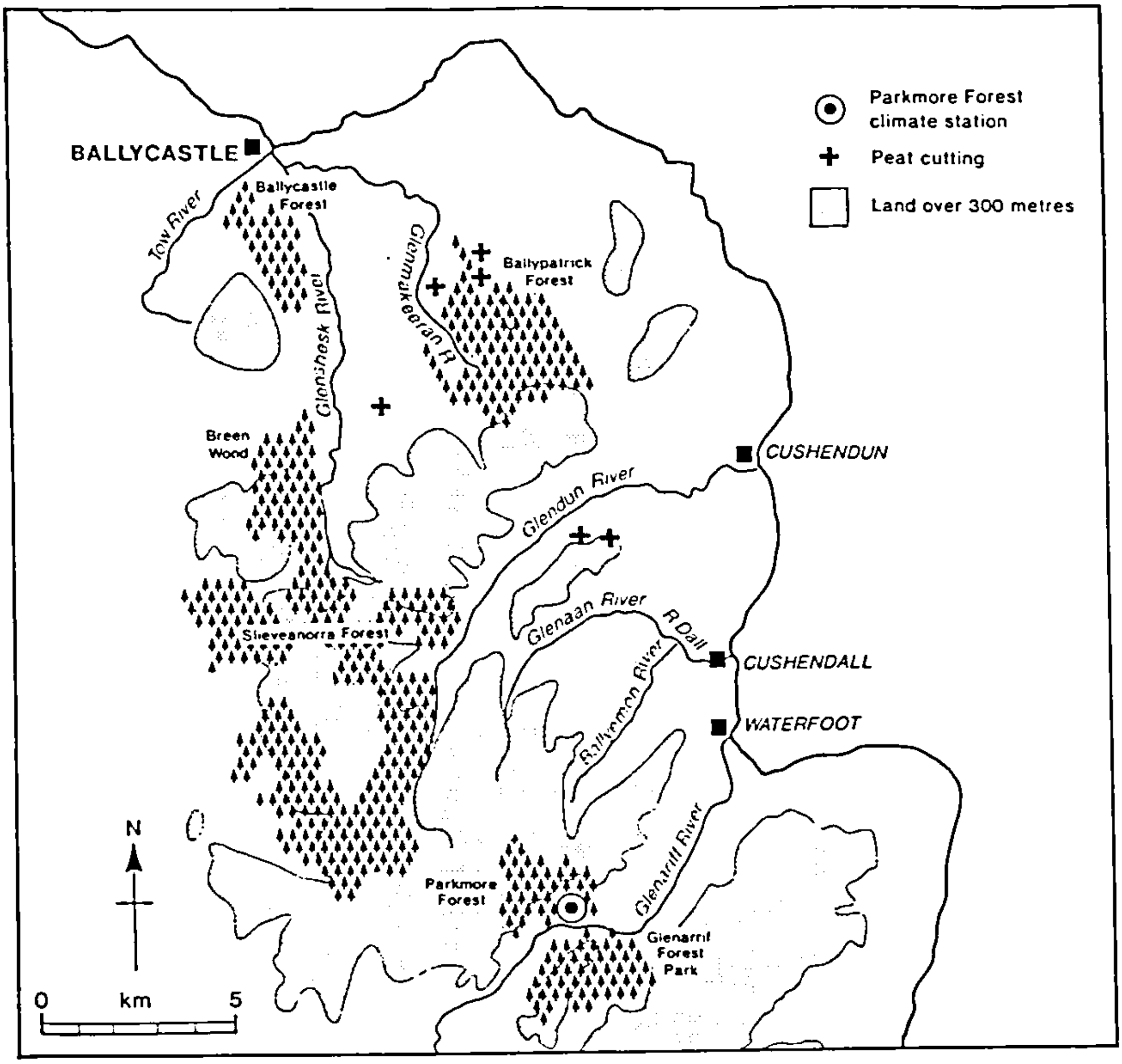

Figure 1 : Location of places in North Antrim mentioned in the text.

especially along the eastern margin (Figure 1). Annual precipitation is generally more than 1250mm, exceeds 1600mm in the highest areas, and between August and January few months are significantly wetter or drier than others. The wet climate has encouraged extensive development of blanket peat, and the upland landscape is devoid of natural tree growth. Drainage takes the form of rivers with relatively small, but steep catchments (River Dall 35 km^2; Glendun River 61 km^2), most of which comprise unforested mountainside. Consequently, discharges display a marked 'peakiness', particularly apparent during periods of high and prolonged rainfall.

Human settlement within the Glens of Antrim consist of isolated farms on lower valley slopes. Each of the major settlements along the coast is located on the valley floor at the point where the glens meet the Irish Sea (Figure 1).

Flood Hazard

Flood hazard assessment requires understanding of a catchment's characteristics in the absence of human activity, responses to climatic change, and how human interference is likely to alter the system.

HISTORICAL PERSPECTIVE

Ungauged catchments and a dispersed settlement pattern prevent analysis of historical precedents for flooding throughout North Antrim, but persistent problems are evident in certain locations.

In Glenariff, despite drainage of the large flood plain comprising marshland in the 1860s, flooding continued to be a recurrent problem. Flood relief schemes in 1880, 1926, and periodically over the last 40 years have failed to alleviate problems associated with the Glenariff River. The flood hazard prevalent at the low-lying coastal settlement of Waterfoot, however, is in part induced by prevailing tidal conditions.

Cushendall lying at the foot of Glenarm and Glenballyemon at the mouth of the River Dall (Figure 1), is prone to flash flooding during heavy rainfall events. To reduce the flood hazard, in 1956 the present defences were constructed providing protection in respect of the 1 in 10 year event. The measures were effective until 1981, since then a number of properties and roads have been subject to repeated flooding as a result of storm runoff from paved areas and buildings being unable to discharge to the Dall due to high river levels.

Historically, Ballycastle has not been prone to flooding, although small-scale inundation in the vicinity of May Street following prolonged or intense rainfall has occurred frequently in recent years (Figure 2).

Smith et al (1987) have commented upon a perceived change in streamflow characteristics of the Glendun River towards a flashier regime, particularly over the last decade. In apparent reaction to this change there has been bed erosion, especially in the channel above Cushendun, sedimentation of Cushendun harbour, and an increased frequency of localised flooding in the village.

It was however, the October 1990 flood which affected much of North Antrim that highlighted the hydrological problems existing in the area.

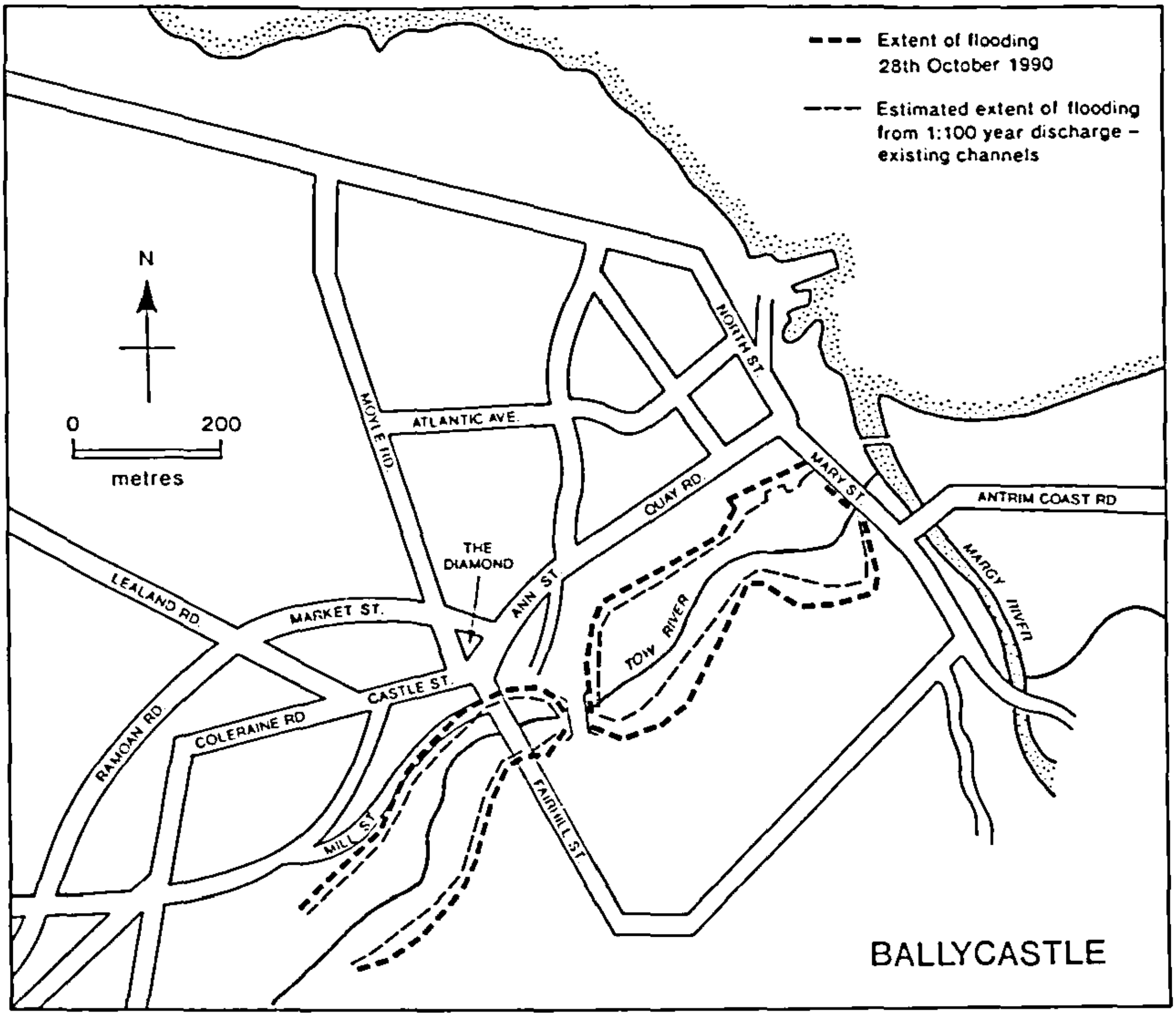

Figure 2 : The extent of flooding in Ballycastle associated with the extreme precipitation event on the 27-28 October 1990.

NORTH ANTRIM FLOOD, OCTOBER 1990

On the night of 27-28 October 1990, a depression with a central pressure of 967mb tracked northeastward across Ireland. Much of County Antrim experienced rainfall in excess of 40mm, but totals of more than 70mm occurred over the Glens of Antrim (Parkmore Forest, 75 mm). Upland falls of 70mm in 24 hours on the Antrim Plateau are certainly noteworthy, but the feature of this synoptic event was that most of the precipitation occurred within a 6-hour period, 0000 to 0600 hours GMT on the 28 October. Such a fall has a return period of once in 270 years, and is categorised as 'very rare' (Natural Environment Research Council, 1975).

With infiltration capacities of river catchments having been reduced by antecedent precipitation over North Antrim amounting to 125 per cent of normal October values, the intense rainfall early on the 28 October gave rise to particularly high flows in several Glens of Antrim rivers.

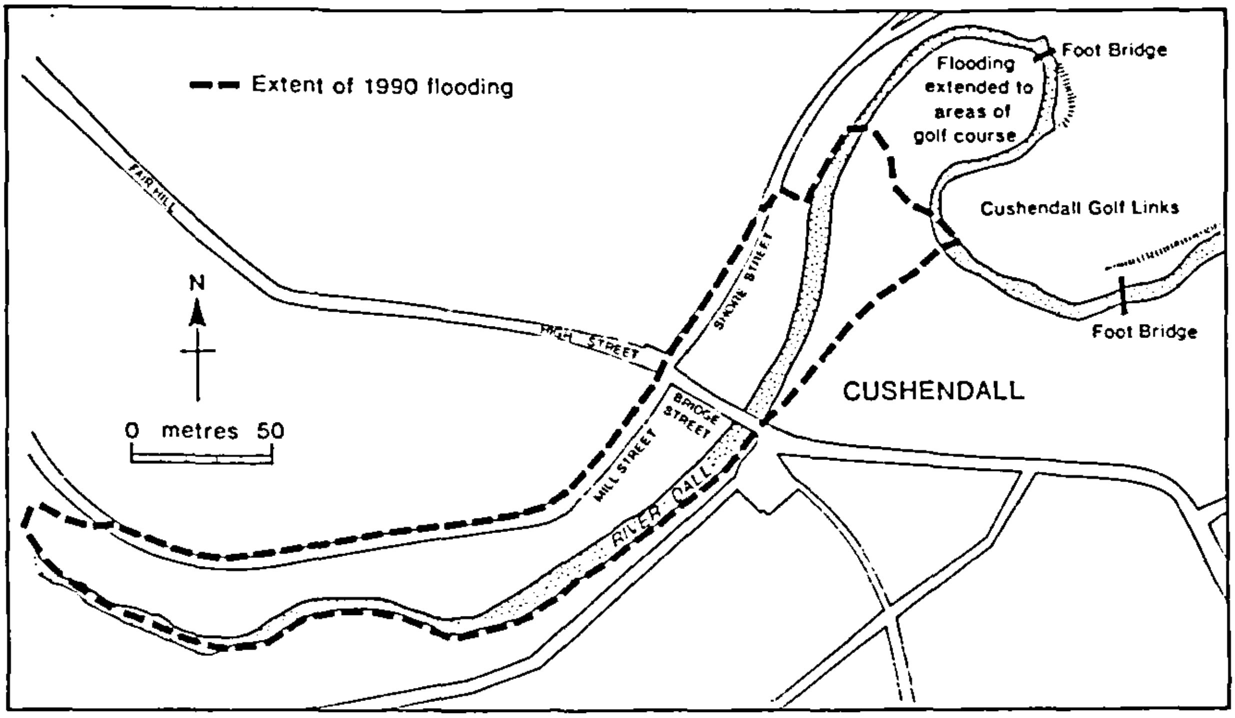

Figure 3 : The extent of flooding in Cushendall associated with the extreme precipitation event on the 27-28 October 1990.

Extensive tracts of agricultural lowland in Glenariff, Glenshesk and along the Glenmakeeran River were inundated.The River Dall was responsible for flooding at Cushendall, and similarly, the River Tow, its catchment partly comprising the slopes of Knocklayd, was primarily responsible for the flooding experienced in Ballycastle.

The extent of floodwater in Ballycastle and Cushendall is depicted, respectively, in Figures 2 and 3. In both settlements many households and business premises were flooded to a depth of 1.5m (Betts, 1992). Clearly, the flooding resulted from the magnitude of rainfall intensities prevailing upon already saturated upper catchments. In the absence of any flow measurement record of the event, calculated flows using catchment characteristics, suggested a discharge of about 50 cumecs was experienced along the Tow River, which relates to an equivalent flood return period of 216 years (Natural Environment Research Council, 1975). The extent of flooding in Ballycastle may have been exacerbated by the narrow span of the Fairhill Bridge, the channel constriction thereby accentuating flow levels along the Tow, upstream of this point (Figure 2). Indeed, more recently, following a wet fortnight in which 125mm of rain fell over North Antrim, a subsequent fall of 15-25mm on the 20-21 December 1991, resulted in the Tow River reaching levels at Fairhill Bridge that again threatened the centre of Ballycastle.

Consequently these events initiated concern to alleviate to some degree the flood threat to Ballycastle and Cushendall. Following a cost-benefit analysis by the Department of

Consequently these events initiated concern to alleviate to some degree the flood threat to Ballycastle and Cushendall. Following a cost-benefit analysis by the Department of Agriculture, structural adjustments to the Tow River commenced in the summer of 1993 and will be completed by the autumn of 1994, providing Ballycastle with protection against the once in 100 year flood event. Similarly, an Environmental Statement was published by the Department of Agriculture in April 1994, presenting proposals for a flood alleviation scheme at Cushendall.

Landuse Change and Flooding

The perceived recent increase of flood occurrences in the Glens of Antrim, appears to have coincided with landuse changes in upper catchments of the area. Blanket peat, up to 3m in depth on gentle slopes, and forming a substantial component of river catchments in North Antrim, has traditionally been used as sheep range and for small-scale peat extraction by hand. During the last 3 decades, however, afforestation, some agricultural reclamation and mechanical harvesting of peat has occurred in these upland catchments. The local community claims that it is these new forms of landuse that have induced change in the hydrological characteristics of the area. The possible role played by each landuse type in causing change will now be placed in perspective. It is important however, to note that the nature of peat composition in Northern Ireland is such that undrained peat seldom exerts the beneficial seasonal effects on streamflow that are often attributed to it. It does not seem to act like a long-term sponge holding back water in winter and releasing it in summer. Residence times in peat, where these have been measured, appear to be quite short, and flash floods are liable to follow closely on heavy precipitation (Hamilton, 1982).

AFFORESTATION

State planting of forests in Northern Ireland has attained a substantial level only during the last 40 years. Government policy that forestry should not compete with agriculture has ensured that blanket peatland has played a major role in recent afforestation schemes (Hamilton, 1982). The rate of planting, however, has been less than anticipated, reflecting difficulties of land acquisition, especially in the better, agriculturally marginal, upland (Tomlinson, 1982). The confinement of forests (mainly Sitka spruce) to peatland and other poor soils, makes forestry extremely difficult, necessitating a shortening of forest rotations.

The wet environment requires a wide area to be drained before planting new ground, and primary drainage takes the form of ploughed furrows, spacing and depth varying according to the need for drainage. A secondary system of cross-drains intercepting water from the primary furrows forms the permanent drainage system of the forest. Concern is therefore understandable, as to the role afforestation of upper catchments may have played in changing flow regimes of rivers in North Antrim. While in the early years the hydrograph on a recently drained catchment will show a more rapid response to rainfall than before ploughing, with tree growth eventually producing a closed canopy, greater interception, increased transpiration, and the infill of drains with branch and leaf litter, the flashiness of the catchment will be reduced. Much of the afforestation in the Glens of Antrim had taken place by the early 1970s. Apart from some planting in Slieveanorra Forest throughout the 1970s, and the small-scale extension of the northwest sector of Ballycastle Forest in the mid 1980s (Figure 1), upland forest comprises mature stands of at least 20 years growth. Consequently, any initial effect of increased runoff due to drainage is no longer a factor although the recent Ballycastle Forest plantings may be partly responsible for enhanced flow levels of the Tow River. In the case of thinning, which is carried out each year, the overall effect on the water yield is hidden. However, clear felling produces an immediate and large increase in the runoff, although the porosity of the surface litter aids infiltration, and to some degree inhibits sheet runoff. At present, any clear felling occurs in relatively small coupes, thereby minimising local changes in runoff pattern.

The maturity of forest stands, and very limited recent planting suggests that the perceived increase in stormflow contribution from the headwaters of Glens of Antrim rivers in recent years, is not the result of afforestation.

LAND DRAINAGE FOR AGRICULTURE IMPROVEMENT

Large areas of hill peatland in North Antrim have traditionally been utilised as rough grazing, but some field drainage has been undertaken as a method of hill farming improvement. This is carried out by the individual farmer although no longer financially assisted by substantial grant-aid, and involves the installation of underdrainage and surface ditches. Such land reclamation for agriculture, however, has been very limited in the Glens, and most improvement occurred during the 1970s. Indeed field drainage was confined mainly to the more fertile lower valley areas, which nevertheless would have enhanced runoff into streams, thereby contributing to increased peak flow during heavy precipitation events.

MECHANISED PEAT EXTRACTION

Historically, by far the most important use of peatland in Northern Ireland is as a fuel. Traditional methods of peat extraction involve turf being hand-cut away from a vertical face which has been drained following cutting in preceding seasons. Any disruption is small-scale, and cut-over portions of peatland regenerate within a few years.

In the early 1980s, mechanised extraction of peat began in North Antrim using small tractor-mounted compact peat harvesters. Partial drainage is often necessary to allow machines onto the peat. This, combined with compaction, and the downslope alignment of extracted cores acting as drainage channels, rapidly directs surface run-off into rivers.

During the past decade, some mechanical peat cutting sites of over 10 ha have occurred in North Antrim, but a multitude of operations involve small patches of little more than 0.5 ha. The headwaters of the Glenshesk River and the upper catchments of the Glenmakeeran and Glendun Rivers, are the main areas of extraction (Figure 1). In Northern Ireland planning permission is required before commercial cutting of peat is begun. No permission is required, however, for the domestic cutting of peat, and locals possessing turbary rights are employing commercial contractors to extract peat on their behalf. The Forest Service has banned mechanical cutting of peat on its land, recognising that damage in the form of waterlogging and reduced stability of the peat renders the land unplantable for future coniferous afforestation.

Concern about machine peat cutting is therefore understandable, but the limited existence of large-scale operations in the study area, prevents placing explicit blame upon this landuse for the increased flashiness of the hydrological regime. Indeed, consideration must also be given to the problem of blanket peat erosion, a feature evident outside the cut areas. It takes the form of gullying, anastomosing channels, hagging, and more rarely, bog flows (Cruickshank and Tomlinson, 1990). The causes are unknown, but the effect further encourages routing of water into the eroding channels, thereby increasing surface runoff.

Climatic Change

A further complication in attempts to accurately estimate the frequency of extreme hydrological events is that climate is known to be far from stable, even on a scale of decades. Climate change reflects fluctuations in the behaviour of the atmospheric circulation . Ireland has experienced enhanced frequencies of westerly and southerly circulations since 1981 although these changes must be viewed only as random fluctuations about the long-term mean (Betts, 1989; Mayes, 1994). Such changes have implications for the geographical distribution of rainfall across the country, and at present an analysis of the frequency of heavy rainfall events over Northern Ireland is being undertaken. Furthermore, much interest exists concerning the possible climatic change scenarios associated with the so-called greenhouse-effect, and particularly future frequencies of extreme climatic events. Great uncertainty exists, however, in the prediction of resulting regional patterns of climatic change (Mitchell et al, 1990), although Rowntree (1990), and Warrick and Barrow (1991) have presented time dependent scenarios of climatic change for the British Isles, including seasonal and annual precipitation change. In considering possible change in terms of extreme climatic events, however, very little can be said about future changes of precipitation intensity and storminess, since much will depend on the regional details of change in the general circulation.

Conclusion

The precipitation event of the 27-28 October 1990 focused attention upon a perceived change in streamflow characteristics towards a flashier regime in North Antrim. An absence of gauge records makes it difficult to ascertain the degree of hydrological change that has taken place, but recent flood alleviation schemes indicate the existence of a worsening flood hazard.

The local community readily asserts that upland landuse change involving agriculture, forestry and peat cutting, is responsible for modifications to the river-flow regime. Examination of the nature of these landuse forms, however, reveals a more complex situation, with peat extraction too readily forwarded as the factor inducing change. Unpublished research, suggests similar hydrological changes may be occurring on the Cuilcagh Mountains, County Fermanagh, where large-scale peat extraction has reputedly raised water-levels in Marble Arch Caves during heavy rainfall events. Unfortunately, no long-term stream gauge records exist for the area, and indeed any specific relationships found between peat cutting and hydrology characteristics in this limestone region, cannot readily be transposed to the markedly different geological area of North Antrim.

It may well be that apart from the large flood event, in the Glens of Antrim, the causal factors of any increase in peak discharges and stormflow may be highly localized, and specific to individual catchments. Clearly, there is a need to monitor the hydrological and landuse changes apparent in these upland catchments in order to calibrate precisely the flood hazard in North Antrim.

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CLIMATE CHANGE AND A NEW WATER BALANCE

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Abstract
The sensitivity of both actual evaporation and catchment runoff to changes in precipitation and potential evaporation are analysed on the basis of the classical formulae for the evaporation ratio (AE/PE) and for a simple soil moisture accounting model. The results of the latter approach suggest the relative sensitivity of runoff to precipitation (Ψ_p) lies within the limits $1 \leq \Psi_p \leq 1 + PE/K_0$ where K_0 is the saturated hydraulic conductivity of the soil. The problems of direct CO_2 enrichment and of upscaling from the pedon and plant scale to catchment scale are briefly mentioned.

Introduction

This meeting is a recognition of the central importance in hydrology of the water balance and its elements. This balance is the expression at catchment or regional scale of the principle of continuity which is the fundamental theorem of hydrology. We have seen in the previous sessions of the meeting that difficulties still remain in determining accurately the distinct elements of the water balance and consequently evaluating the acceptability for different purposes of the residual out-of-balance term.

This problem becomes even more complicated if climate change has to be taken into account. Global climate models (GCMs) include in their outputs estimates of precipitation, potential evapotranspiration and soil moisture for prescribed levels of concentration of CO_2 and other greenhouse gases. The linking of climate models and hydrologic models to predict the impact of climate change on hydrology and water resources is still at a very early stage. Thus, Chahine (1992) writes:

> "Climate models have yet to be developed which account for the full hydrological cycle and the interactions with the atmosphere, oceans and land".

The present paper concentrates on discussing in simple terms some of the important principles involved in the problem rather than exploring the jungle of multi-parameter climate models and multi-parameter hydrologic models.

Long-term Equilibrium

The problems of predicting changes in hydrology and in vegetation arising from climate change can be analysed initially on the basis of perturbations from a condition of long-term equilibrium. Under the principle of geographical zonality (Marsh 1864, Budyko 1950), it may be assumed that the key water balance element of actual evapotranspiration at the biome scale would depend only on the long-term average of the precipitation and potential evapotranspiration, the biome vegetation (at temporary or at climax equilibrium), and the average soil conditions.

The simplest assumption that can be made is that, at the time scale of interest, the soil conditions and the vegetation conditions are determined by the long-term averages of the basic climate variables of precipitation and potential evapotranspiration. Under this restrictive assumption, actual evapotranspiration (AE) can be expressed as a function of precipitation (P) and potential evapotranspiration (PE) and can be written in the form:

$$\frac{AE}{PE} = \phi\left(\frac{PE}{P}\right) \qquad (1)$$

where PE/P is the aridity index used by Budyko (1950) to characterise the type of biome appropriate to a given climate. The author thinks it appropriate to refer to equation (1) as the Budyko hypothesis.

The classical empirical formulae used in climatology to estimate actual evapotranspiration (Schreiber 1904, Ol'dekop 1911, Turc 1954) can all be expressed in the form of equation (1). The family of curves defined by the differential equation of Bagrov (1953):

$$\frac{dAE}{dP} = 1 - \left(\frac{AE}{PE}\right)^n \qquad (2)$$

can be expressed in the form

$$\frac{AE}{PE} = \phi\left(\frac{PE}{P}, n\right) \qquad (3)$$

The solution of equation (2) closely approximates Schreiber's curve for the parameter n = 1 and corresponds exactly to Ol'dekop's curve for the parameter value n = 2. Budyko and Zubenok (1961) examined the long-term water balance for 1200 regions throughout the USSR and found that the data were all within the limits defined by the Schreiber curve and the Ol'dekop curve.

Recent work (Rodrigues-Iturbe et al. 1991) indicates that the reasonable assumption that a fixed proportion of local evaporation is re-precipitated locally may lead to a bi-modal probability distribution of soil moisture at higher levels of input variance. Above a threshold, the distribution in their example has two modes - the larger at about 25% of the long-term mean and a smaller one at about 80% above that mean. This offers one explanation of the persistence of runs of dry years and wet years. In the following discussion, however, the analysis will be made on the basis of perturbation from a single equilibrium condition.

Sensitivity to change in P or PE

All of the classical empirical formulae approach the asymptote

$$AE \quad = \quad P \qquad (4)$$

as the value of P approaches zero (i.e. for extreme aridity) and are asymptotic to the limit:

$$AE \quad = \quad PE \qquad (5)$$

as the value of P increases without limit (i.e. for extreme humidity). Accordingly, the value of ϕ (PE/P) monotonically decreases from unity to P/PE as the aridity index (PE/P) increases from zero to infinity.

For any relationship which conforms to the Budyko hypothesis represented by equation (1), the sensitivity of actual evapotranspiration to precipitation and the sensitivity to potential evapotranspiration can be shown to be related. We can write the sensitivity as

$$\frac{\Delta AE}{AE} \quad = \quad (\sigma_p)\,\frac{\Delta P}{P} + (1-\sigma_p)\,\frac{\Delta PE}{PE} \qquad (6)$$

where σ_p can be written in terms of the function ϕ (PE/P) in equation (1) by

$$\sigma_P = - \frac{\phi'(PE/P) . (PE/P)}{\phi(PE/P)} \qquad (7)$$

The above relationships are valid for all cases obeying the Budyko hypothesis of equation (1). The value of the sensitivity factor will be σ_P = 0 for a low aridity index (i.e. for tundra) and σ_P = 1.0 for a very high aridity index (i.e. for desert). The intermediate values depend on the form of the function used in equation (1).

Since the long term catchment runoff is the difference between precipitation and actual evapotranspiration, the relative sensitivity of catchment runoff to relative change in precipitation or relative change in potential evapotranspiration can be similarly evaluated (Dooge 1991a, 1991b). For catchment runoff the relevant sensitivity factor Ψ_P is always greater than 1.0 in contrast with the above sensitivity factor σ_P which is always less than 1.0 . For all the empirical equations considered above the value of Ψ_P approaches 1.0 as the aridity index becomes vanishingly small. For the Ol'dekop and Turc-Pike formulation, the value of Ψ_P approaches 3.0 as the aridity index (PE/P) increases without limit. However for the Schreiber formulation, the value of Ψ_P increases without limit as the aridity index increases. For the Bagrov differential equation the value of the sensitivity index increases from Ψ_P=1 at PE/P = 0 to Ψ_P=n+1 at PE/P= ∞.

Simple Soil Moisture Accounting

A second reasonable step in studying the sensitivity of actual evapotranspiration or of catchment runoff to climatic factors is to apply some simple form of soil moisture accounting. The simplest form of the basic water balance of soil moisture is obtained when it is assumed that the conditions are dry enough for the soil never to become saturated so that infiltration is always equal to precipitation and that evaporation is always less than the potential rate. For these assumptions the soil water balance is given by

$$\frac{dW}{dt} = P(t) - AE(t) - Q_b(t) \qquad (8)$$

where $W(t)$ is the volume of soil moisture storage, $P(t)$ is the rate of precipitation, $AE(t)$ is the rate of actual evapotranspiration, and $Q_b(t)$ the rate of base flow fed into the drainage system from the subsurface storage.

The soil moisture balance given by equation (11) can be solved analytically by assuming that both $AE(t)$ and $Q_b(t)$ are proportional to the $W(t)$. The first of these assumptions

$$\frac{AE(t)}{PE(t)} = \frac{W(t)}{W_{max}} \qquad (9a)$$

or

$$AE(t) = a.W(t) \qquad (9b)$$

is that of the Budyko bucket used in the first generation of climate models to incorporate some representation of the hydrological cycle (e.g. Manabe 1969). The second assumption

$$Q_b(t) = b.\,W(t) \qquad (10)$$

is the basic assumption underlying an exponential form of the base flow recession curve which has been widely used in applied hydrology.

This approach has been applied to the sensitivity of catchment runoff in the continental United States by Schaake (Schaake and Liu 1989, Schaake 1990) and independently by the present author to the case of two-season climate with varying lengths of the rainy season (Dooge 1991a, 1991b). In these studies the sensitivity of runoff to precipitation and potential evaporation was expressed as

$$\frac{\Delta Q}{Q} = \psi_P \cdot \frac{\Delta P}{P} + \psi_{PE} \cdot \frac{\Delta PE}{PE} \qquad (11)$$

where analogous to equation (6) above we have

$$\psi_P + \psi_{PE} = 1 \qquad (12)$$

and where ψ_P is given by

$$\psi_P \quad = \quad \frac{1 + \phi'(PE/P) \,.\, (PE/P)^2}{1 - (PE/P) \;.\, \phi(PE/P)} \qquad (13)$$

where ϕ (PE/P) is again the function in equation (1) defining the Budyko hypothesis.

For very low values of the aridity index (i.e. tundra) AE is asymptotic to PE and in the limit we have

$$\psi_P \quad \approx \quad \frac{1}{1 - PE/P} \qquad (14)$$

which is equal to 2.0 for PE/P = 0.5 and approaches 1.0 as this aridity index decreases to zero. For high values of aridity index, the assumption that the soil never becomes saturated is satisfied and the simplified form of equation (8) can be readily solved analytically for the condition of seasonal variation but no inter-annual variation. For these conditions it can be shown from the average annual water balance that

$$\frac{AE}{PE} \quad = \quad (\frac{a}{a+b}) \frac{P}{PE} \qquad (15)$$

Substitution of equation (15) into equation (13) gives us

$$\psi_P \text{ (large PE/P)} \quad = \quad 1 \qquad 16)$$

This result is significant as the form of the classical empirical formulae give values of ψ_P ranging from 3.0 for Ol'dekop to infinity for Schreiber (Dooge 1991a).

Further analysis reveals that the maximum value of the sensitivity factor ψ_P occurs in this simple model close to the point of discontinuity where the aridity index is such that the soil just reaches saturation for a very short period each year. This suggests that the most sensitive areas may not be in the most arid regions but may be close to the ecotone between forest and steppe with an aridity index PE/P around unity. Analysis suggests (Dooge 1991a) that the value of the sensitivity index would be bounded by

$$\psi_p(\max) \leq \frac{a + b}{b} \qquad (17)$$

where a and b are the parameters defined by equations (9) and (10). It is important to note that if the hydrologic component of the climate model does not allow for subsurface drainage (which is the case for many GCMs) then $b = 0$ in equation (10) and the bound in equation (17) becomes infinity.

Carbon Dioxide and Vegetation

If we are interested in the effect on the water balance of a marked increase in greenhouse gases in the atmosphere, the effect of the new level of CO_2 on vegetation cannot be ignored. The response of plants and ecosystems to carbon dioxide enrichment can be grouped in terms of (a) primary biochemical and cellular responses (including stomatal aperture), (b) secondary biochemical, cellular and whole plant responses (including plant water status, (c) tertiary whole plant responses (including leaf area), (d) ecological level responses (including plant-plant competition and symbiosis), and (e) genetic responses (including adaptions).

The processes involved are all highly interactive and the total effect involves a large number of feedbacks both positive and negative. As the carbon dioxide content of the atmosphere is increased, the stomatal pores of the leaves tend to close because of the increased gradient through the leaf surface of the concentration of carbon dioxide. The magnitude of the resulting reduction in conductance has been estimated for a doubling of the CO_2 concentration from 330 ppm to 660 ppm as being between 34% and 40%. The reduction in transpiration will be less than the reduction in conductance due to the rise in leaf temperature and other possible feedbacks.

The effect of differing assumptions concerning the direct enrichment effect of carbon dioxide on vegetation can best be exemplified in the case of semi-arid or arid catchments in which the evaporation process is water-limited rather than energy-limited. Revelle and Waggoner (1983) used the empirical relationship due to Langbein et al. (1949) to investigate the effect of climate change on 12 basins in Arizona. They estimated that for a particular scenario the runoff for the five wettest of these basins would decrease by 41%. Idso and Brazel (1984) repeated the study of Revelle and Waggoner but assumed that due to stomatal changes the evapotranspiration over the vegetated part of the catchment (which varied from 24% to 61%) would be reduced to two-thirds of the original rate assumed by Revelle and Waggoner. When this change was made, the effect of the CO_2 doubling scenario used by Revelle and Waggoner is changed from a decrease in runoff of 41% to an increase in runoff of 42%. It is not a wise use of research resources to continue studying more complex models of catchment response before dealing with the question of this disparity.

Woodward (1987), in an interesting study, compared (a) the relative changes in the stomatal density of leaves under different concentrations of carbon dioxide in the laboratory with (b) the relative changes in the stomatal density of dried leaves preserved in the university herbarium at various times over the past 200 years. The average rate of reduction in the preserved historical plants was found to be about 60% of the rate of reduction for similar increases in carbon dioxide concentration under controlled laboratory conditions.

As in the case of hydrologic modeling, one of the key questions in all these studies is that of scale and of the relationship between results obtained at a leaf or plant scale and those at a regional and a biome scale. The possibility of scaling up from the single stoma to the single leaf, from the single leaf to the individual plant, from the plant to an extensive canopy and ultimately to a region, has been well reviewed by Jarvis and McNaughton (1986). As mentioned earlier, much work remains to be done, but research in this field is essential if the impact of climate change on the vital area of water resources is to be properly evaluated. It obviously requires a dialogue between hydrologic modelers, climate modelers and modelers of vegetative systems of all types.

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ABSTRACTS OF
SHORT PRESENTATIONS

1. THE DYNAMICS OF WATER BALANCE FOR DIFFERENT SOILS IN SLOVENIA FROM 1961 - 1990

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The dynamics of soil water balance at 18 locations in Slovenia from 1961 to 1990 were studied and also the comparison with the extremely dry year of 1992 was made. The monthly values of potential evapotranspiration, the excess water and the deficit of water in two types of soil (the first one with 60 mm and the second one with 100 mm water content in the upper layer) were calculated according to Thornthwaite's method. The analysis of water balance for the entire observation period from 1851 to 1993 was made for Ljubljana. The potential evapotranspiration and the deficit of water increased, but the excess water declined in the investigated period. The same is true for the last 30 years for nearly all locations studied in Slovenia.

2. WATER BALANCE OF THE NORE RIVER BASIN

Eugene P. Daly

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The Nore River Basin is situated in the southeastern part of the Central Plain of Ireland and covers an area of 2,530 km^2. The Geological Survey carried out an extensive hydrogeological investigation in the catchment between 1975 and 1981.

The study included the calculation of an annual water balance for the ten year period between 1972 and 1981. Monthly values were used for the inputs and outputs.

Values for rainfall and potential evapotranspiration were obtained from the Meteorological Service and those for runoff were provided by the Office of Public Works. The Aslyng scale was used to estimate the actual soil moisture deficit from which values for actual evapotranspiration were obtained. Estimates of changes in groundwater and soil moisture storage are based on the water level hydrographs of a number of wells and the soil moisture conditions. Estimates were made of a number of unmeasured variables in the system which suggested a small net outflow.

The results show that the balance achieved is quite good with the level of error being relatively small. However, it is most likely that larger errors in the individual components are being averaged out in the water balance equation.

3. APPLICATION OF MONTHLY WATER BALANCE MODELS TO IRISH CATCHMENTS

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Water balance models allow the quantitative evaluation of the distribution of total water among the various components in a basin based on available hydrological and meteorological data (*i.e.* precipitation, and possibly potential evapotranspiration, temperature, humidity and streamflow).

The primary application of these models are to aid water resource management decisions concerning water supply/availability (drinking water, adequate dilution for pollutant discharges, hydroelectricity, *etc.*). Specifically, applications are the real time forecasting of catchment yields, the generation of flow records for ungauged periods based on available precipitation and evaporation records and the preliminary investigation of other ungauged catchments having similar geographic characteristics. By building up experience in the behaviour of the model parameters to geographic characteristics, effective models can be developed for ungauged catchments which will provide reasonable predictions in terms of the distribution of rainfall among the other hydrological components.

The model formulation comprises a deterministic rainfall-runoff filter which computes monthly streamflow for monthly inputs of precipitation and potential evapotranspiration and a random deviation function which accounts for the "un-observed" factors, such as measurement errors, Thiessen errors and model imperfections. The limitations of the model are that its applicability is limited to basins whose areas are less than 5,000 km^2 (as it only uses one storage component), without significant snow and frost phenomena and without important natural or artificial lakes. No surface water storage is considered by the model, consequently transmission time via lake systems or snow melt must be less than 1 month.

This study demonstrates the usefulness of a simple monthly water balance model for predicting basin storage and streamflow components. The model was applied to three basins, two with relatively flashy streamflow characteristics, namely the Nore and Suir basins and one having a damped lowland characteristic, namely, the Brosna basin. The model streamflow predictions were found to agree well with observed flow proving the model as an effective tool for predicting monthly streamflow. Furthermore, these simple models exhibited reasonable capability in distinguishing slow and fast flow components. The study also highlights the potential for application of the model to ungauged basins of similar geological circumstance.

The effect of data inaccuracies were investigated using the Suir basin model. The study revealed that if precipitation was adjusted by +5% for systematic error then to achieve a reasonable water balance the potential evapotranspiration had to be subsequently increased by 15%. This indicates that if precipitation is underestimated, as is the general belief among hydrologists, then Penman potential evapotranspiration is also under-estimated. Model Analyses of the Nore basin also revealed this.

A series of adjustments to potential evapotranspiration were carried out to assess the influence of errors in evapotranspiration on the model efficiency R^2. The findings revealed that over-estimating potential evaporation has less impact on model results than under-estimating it. The converse is true regarding the precipitation data. Therefore, it is suggested that a fundamental requirement for model application is that the basin discharge plus potential evapotranspiration is greater than precipitation, assuming that such data is available.

4. PREPARATION OF A WATER BALANCE FOR FOUR CATCHMENTS

Mícheál Mac Cárthaigh

Water Resources Section, Environmental Protection Agency

Water balances were prepared for four catchments: The Glyde, Dee, Finn and Blackwater (Monaghan) Catchments for the period December 1975 - November 1977 as part of the North-East Groundwater Investigation.

The catchment area to the hydrometric gauging station in each catchment was determined with results as follows: Glyde 270 km^2; Dee 307 km^2; Finn 175 km^2; and Blackwater (Monaghan) 126 km^2.

Estimates of Potential Evapotranspiration (P.E.), based on the Penman Formula, were used to estimate the Actual Evapotranspiration, using the Aslyng Scale to relate P.E. to A.E..

For each catchment, the runoff hydrograph was separated into groundwater and surface runoff.

Estimates of the groundwater inflow, outflow and change in storage were calculated for each month of the study.

The study showed that the elements of the hydrological cycle could be balanced within experimental error in the period December 1975 - November 1977; groundwater infiltration took place in winter/spring; there was no substantial net change in storage over the period examined in any of the four catchments; groundwater runoff was roughly similar in the Finn and Blackwater (Monaghan), at 13% and 15% of total rainfall (respectively), and groundwater runoff in the Glyde and Dee was 24% of the total rainfall (these estimates can be altered substantially depending on where the groundwater separation line is judged to be).

5. THE GROUNDWATER COMPONENT OF THE NORE RIVER SYSTEM

Eugene P. Daly

Groundwater Section, Geological Survey of Ireland

The Nore River Basin is situated in the southeastern part of the Central Plain of Ireland and covers an area of 2,530 km^2. The Geological Survey carried out an extensive hydrogeological investigation in the catchment between 1975 and 1981.

Baseflow measurements throughout the river system proved to be a very useful tool in determining the groundwater flow regime in this river basin.

A large number of lowflow measurements, mainly by the Hydrometric Section of the Office of Public Works, were made throughout the river system in late August/early September of 1976, at the end of a long dry summer. Measurements were also made in other years on particular parts of the system.

An initial analysis showed that the major groundwater contributions to baseflow in the river system occur in the areas upstream of the stations at Brownsbarn Bridge (Inistioge) and Johns Bridge (Kilkenny City) with smaller contributions from the area upstream of Durrow (Co. Laois). The analysis of lowflow data provided an initial conceptual hydrogeological model for use during the earlier part of the investigation and gave the locations of the discharge areas of the main aquifers.

The baseflow analysis also suggested interpretations for some hydrogeological features that could not be explained early on in the study or confirmed conclusions arrived at by other means.

The groundwater component of riverflow at the Brownsbarn Bridge gauging station, for the period 1972-1981, was estimated using a hydrograph separation techique in combination with a number of well hydrographs. The mean groundwater component of this river system is about 250 mm y^{-1} over the river basin, 26% of rainfall or 50% of total run-off.

6. MODELLING AND FORECASTING OVERLAND FLOW

Natasha Ariff[1], Tim Gleeson[2] and Jim Kiely[3]

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A full computer working model of a previously described simple conceptual single reservoir hydrological model of the occurence of overland run-off was developed. Using only rainfall data, a constant percolation rate for the soil, and estimated soil moisture deficits, the actual overland run-off from the measured data from trials was compared to that predicted by the model.

Good correlation was obtained between measured and observed data for all events in winter and for larger rainfall events in summer. The model needs further refining using smaller time intervals and a better estimate of soil moisture to predict minor overland run-off events in dry summer weather. As well as being useful for forecasting overland run-off risks, the model has potential also for forecasting difficult surface conditions for intensive grazing.

7. MODELLING VARIABILITY IN HYDROLOGIC INTERACTIONS AT THE LAND SURFACE/ATMOSPHERE INTERFACE

Dr. M. Bruen

Centre for Water Resources Research, Department of Civil Engineering, University College Dublin

The CWRR has been coordinating an EU funded research contract investigating the modelling of moisture fluxes at the land surface/atmosphere boundary. The aim was to improve the land surface parametrization used in Global Circulation Models. The CWRR team developed a simple and computationally fast method for modelling moisture fluxes through a bare soil column. This model was subsequently expanded to include the effects of variability in rainfall and soil properties and this poster reports some of the results obtained with it.

Monte Carlo simulation using a large number (500) of soil columns was used to investigate the sensitivity of the single column code to variability in rainfall and in soil properties. Dealing with rainfall variability required two stochastic components; the first determined whether a particular soil column received any of the precipitation specified for a particular time interval. If it did, then the amount of precipitation it received was determined by the second stochastic component. Both random components used probability distributions with parameters chosen so that the total rainfall amount applied to all columns matched a given forcing climate specified at half-hourly intervals. In most cases a total of two years was simulated but results have reported only for the second year.

Variability in soil properties was included by using a scaling parameter which governs the hydraulic properties of the material. The current version of the model allows this to be included by itself or together with rainfall variability. Two types of soil, a loam and a sandy soil are modelled and some typical results are presented. For the loam soil the annual amounts of evaporation, runoff and drainage are not affected very much when soil variability is taken into account, even though there is considerable variation between individual soil columns. However the annual amounts are influenced much more by rainfall variability. The reverse is true for the sandy soil.

The effects of the depth of the soil column and the assumption of either free drainage or a water table as a lower boundary condition are discussed.

8. EVALUATION OF AREA EVAPOTRANSPIRATION BY MULTISENSOR DATA

Andreas Weimann, Maria V. Schönermark, Helge Witt

DLR - German Aerospace Research Establishment, Institute of Space Sensor Technology

Satellites are giving us the possibility to get information from extended and distant places. By using multisensor data we are trying to estimate the evapotranspiration using satellite data. A model has been developed which needs satellite data from visible, thermal and microwave parts of the spectrum. At the moment, there exist different types of satellites with channels situated in these regions. The model takes into account different types of vegetation. With the shortwave (VIS) and nearinfrared (NIR) data the land use is classified with regard to it's behaviour to evapotranspiration. So every class represents a hydrological class. The thermal channels give us information about the temperature of soil and vegetation. With these parameters, as well as the radiation flux, the potential evapotranspiration can be estimated and afterwards a specific evapotranspiration of every hydrological class can be determined. With the soil moisture, obtained by microwave remote sensing, the actual evapotranspiration can be obtained. Taking into consideration the size of the land use patches, the area evapotranspiration can be estimated.

First, analyses have been done with TM satellite-data. An atmospheric correction method has been developed and has been used with the TM-data with success. The corrected images have been classified to obtain the hydrological classes.

For the soil moisture estimation, the P- or L-band would be ideal, because these wavelengths penetrate deep enough into the soil in order to receive information about the soil water conditions. But at the moment, such data are not available. As a result we have taken the data of ERS-1 satellite (C-band) to investigate the relationship between satellite-signal and soil moisture.

9. WATER BALANCE AND INLAND FISHERIES

Christopher Moriarty

Fisheries Research Centre, Abbotstown, Dublin

The salmon is the key species of fish in Irish inland waters. Apart from its popularity in recreational fishing, its survival is an essential factor in the existence of many rural communities. It is also the species most sensitive to variations in the quantity and quality of surface water. If salmon can thrive in a river system, so can all other species.

The salmon needs:

- continuous dry-weather flow in rivers
- dissolved oxygen close to saturation level
- pH above critical level of 5.5
- excellent water quality at all times.

10. THE EFFECT OF SOIL MOISTURE DEFICITS ON THE GRAIN YIELD OF SPRING BARLEY

Michael J. Conroy

Teagasc, Oak Park Research Centre, Carlow

Nine experiments were carried out over the 3-year period 1988 - 1990 to test the effect of cultivar on the grain yield and quality of spring-sown malting barley (Conry, 1994). The experiments were conducted in the south-east of Ireland on three different soil types which were representative of the range of soils devoted to malting barley growing in Ireland.

There was large seasonal variation in grain yield which was attributed largely to the impact of the weather pattern on the different soil types. It was difficult to relate grain yield to the recorded weather parameters (mean daily temperature, sunshine, radiation, rainfall) except to rainfall in 1990.

The results of these experiments show that soil moisture deficits do in fact reduce spring barley yields and the effect depends largely on soil type and the crop growth stage at which the soil moisture deficit occurs.

Reference:

Conry, M.J. 1994. Comparative effect of six cultivars at four rates of nitrogen on the grain yield and grain quality of spring-sown malting barley in Ireland. *Journal of Agricultural Science*, Cambridge, 122, 343-350.

11. IRRIGATION SCHEDULING IN NORWAY

T.H. Sivertsen

Department of Plant Pathology, Norwegian Plant Protection Institute

The Norwegian Plant Protection Institute is responsible for the agrometeorological services of the country. The agrometeorological service includes irrigation scheduling for the farmers, and the Plant Protection Institute gets the necessary weather data for making warnings and recommendations for protection of plants against pests and diseases. The agrometeorological service is running 38 automatic weather stations placed in rural districts all over the country. It is a close collaboration with the national meteorological institute that provides the necessary weather forecasts on a local scale. The irrigation scheduling will be field specific. The farmers may access the automatic service on a 24 hour's basis by means of telephone or PC equipped with modems.

12. MODELLED AND MEASURED WATER FLUXES IN A FOREST ECOSYSTEM

J.J. Aherne, T. Cummins and E.P. Farrell

Forest Ecosystem Research Group, University College Dublin

The Forest Ecosystem Research Group (FERG) control a series of intensive forest monitoring plots with the aim of quantifying the effects of atmospheric deposition on forest ecosystems. A knowledge of water fluxes through the ecosystem can greatly contribute to the understanding of biogeochemical processes and acidification. Precipitation and throughfall can be measured directly, however, fluxes such as soil water and evapotranspiration cannot. Computer simulation models can be used to estimate these fluxes. The initial study, which is presented here, describes four years of measured and simulated water fluxes for Ballyhoolly, Co. Cork.

13. THE PEAT INDUSTRY IN THE IRISH CLIMATE

Donal Wynne

Civil Engineering and Environmental Control, Bord na Móna

Bord na Móna produces milled peat mainly for Electricity Generation, Briquetting for industrial and domestic fuels and moss peat for Horticulture. The peat is milled at about 85% Moisture Content and air dried to about 45% Moisture Content before harvesting. The process is therefore totally weather dependent. Typically in Ireland it takes 3 days of anticyclonic weather in June/July to produce a harvest. Twelve such harvests are required to achieve targets.

Water is drained from the bogs into the local streams and rivers. Peat silt can be carried into the rivers unless precautions are taken. The silt can settle out at bends in rivers and at the confluence of streams and rivers and can cause nuisance to fishermen, farmers and other river users. Silt control ponds are installed on all outfalls to trap and store the silt. They are designed to suit the catchment area, the predicted run-off and the maintenance requirements.

This poster looks briefly at how the peat industry interacts with the natural environment; how it deals with and is affected by the constantly changing and unpredictable Irish Climate.

AGMET Publications

First Report	1985
The AGMET Index : Irish Scientists Concerned with Agricultural Meteorology	1986
Weather and Agriculture (booklet in association with Agricultural Credit Corporation plc (ACC))	1986
Climate, Weather and Irish Agriculture (Gen Ed; T. Keane) (textbook sponsored by ACC; available from Dr J Collins, Faculty of Agriculture, UCD, Belfield, Dublin 4)	1986
Automatic Climatological Recording (Chairman; W Burke) (Report prepared by AGMET Subgroup on Automatic Weather Recording)	1987
Proceedings of Conference on Weather and Agriculture	1988
Irish Farming, Weather and Environment (Tom Keane, Editor) (sponsored by Teagasc; available from Teagasc, 19 Sandymount Ave, Dublin 4)	1992
Weather, Soils and Pollution from Agriculture (Compiled by Marie Sherwood) (available from AGMET, c/o Meteorological Service, Glasnevin Hill, Dublin 9)	1992
Future of Irish Agriculture - Role of Climate : Proceedings of 1992 AGMET Conference (Editor; J F Collins) :	1993
The Balance of Water - Present and Future : Proceedings of AGMET/ Royal Met. Soc. Conference 1994 (General Editors : T. Keane and E. Daly) (available from AGMET Group c/o Meteorological Service, Glasnevin Hill, Dublin 9)	1994

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