The Influence of Advection
on Water Quality Variation
in a Deep Australian Impoundment

by

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Submitted in fulfilment
of the requirements for the degree of
Doctor of Philosophy

UNIVERSITY OF TASMANIA
HOBART

graduating 1986

June 1985

DECLARATION

I declare that except as stated herein this thesis contains no material which has been accepted for the award of any other degree or diploma in any University, and that to the best of my knowledge and belief this thesis contains no copy or paraphrase of material previously published or written by another person, except when due reference is made in the text.

JM Ferris (21.6.85)

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ABSTRACT

Data spanning twenty years (1961-1980), for a site adjacent to the dam wall in Lake Burragorang (34° 55' S; New South Wales, Australia), is analysed with the emphasis placed on physical and chemical (water quality) records. Lake Burragorang is a deep (105 m), dendritic (shoreline development 11.53), essentially warm-monomictic lake with a median bulk retention time of 1.8 years. Inflows are unpredictable seasonally and annually, however, and the study included years with retention times ranging from 0.5 - > 22.0 years.

The aim of the investigation was to assess the long-term behavioural variation and to explain the major factors contributing to that variation.

The cycles of thermal and oxygen stratification are profoundly influenced by inflow/outflow relations. The monomictic cycle is interrupted in c. 50% of years, by cold underflows which help to prevent the short period of circulation (1 - 2 months) which occurs in other years. A significant role is almost certainly played by the mid-depth outflow current, which is presumed to interfere with convective mixing deeper than c. 44.5 m (below Full Supply Level). The sub-surface offtake is not used in dry years.

A seasonal cycle of inflow is found for the lake, and the interchange between interflows and underflow is predicted with about 80% success on the basis of 4 temperature measurements, from the two major inflowing rivers and the surface and bottom adjacent to the dam.

A comparison of monthly mean profiles of temperature and dissolved oxygen from wet and dry year groups (differing in total annual inflow by approximately one order of magnitude) reveals that advective processes increase the summer heat income and help to distribute heat more deeply into the water column. Subsequently, these processes increase the rate of autumnal heat loss, yielding an almost equal winter heat content and effectively increasing the annual heat budget by 22% of the dry year budget. The downward movement of heat is found to be most closely related to the volume of water subtracted through the sub-surface offtake, rather than to

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the volume of influent water. Within the accuracy of profiling (6 m intervals) the wet and dry years appear not to differ with respect to the depth of the convectively mixed layer during the cooling phase (March - June). The summer maximum of Schmidt stability is lowered by only 1% in wet years, but autumnal stability is lowered by up to 33% while, in winter, stability is enhanced by > 100% of the dry year stability. The maximum Birgean wind-work increases by 100% in wet years as a result of advective effects which are outside its conceptual framework. The volumetric hypolimnetic oxygen depletion rate is doubled in wet years.

Turbidity is linearly related to the volume of underflows, but the relationship is much weaker with respect to interflows which generally occasion lower levels of turbidity (for a given inflow volume) at the outflow site.

Lake Burragorang is phosphorus limited and a close relationship is found between total phosphorus and chlorophyll concentrations.

Inflow and outflow are considered the major contributors to the variation in Lake Burragorang's behaviour for the period 1961 - 1980, altering thermal and oxygen stratification behaviour, markedly affecting water quality, and introducing the nutrients required to support algal growth above the usually low levels of dry years (1 - 5 mg m^{-3} of Chlorophyll-a). No obvious period of trophic upsurge is found for the lake, and the data presented does not indicate any significant trend towards eutrophication, though such trends will be difficult to determine against the advection induced variability of the system.

Acknowledgement

There are many people who have contributed, and in many ways, to the completion of this work. The Sydney Metropolitan Water, Sewerage and Drainage Board provided both the lake and access to the data accumulated over 20 years of monitoring it. I also gratefully acknowledge the Board's financial assistance. I would particularly like to thank the staff of the Biological Section (under Ian Smalls) for their help and friendship, and in the case of both Ian Smalls and Colin Heath for having me stay at their homes. I also thank the members of the Chemical Section (under Peter Sadler) and the people at Warragamba for their assistance in the laboratory and on the boats. The task of accumulating some quite old records took me to many other sections within the Water Board, and I am grateful for the help extended to me by all those who answered questions and helped provide the raw material for this work, especially those who replied to my written request for early inflow records with a small mountain of photocopies.

I am grateful to my supervisor, Peter Tyler, for laying the groundwork and then entrusting the project to me to take in directions only partly envisaged at the outset. I also thank him sincerely for being there when I most needed his experience, towards the end, and for many friendly curries throughout. I also thank Professor W. D. Williams for presiding over my transportation to Tasmania in the first place.

I am deeply indebted to Trevor Olesen, Jenny Read, Jocelyne Hughes and William Grove for the help with figures, typing, photocopying and general moral support, and to Tim Walker for the means of word-processing and presentation. Yvonne Shaw provided cheerful help with typing, for which I am also grateful. My computing and statistics were greatly assisted by both Peter Minchin and Brad Potts.

Less tangible but important help came from friends at the University of Tasmania Botany Department who assisted a "mainlander" to settle into Hobart and to feel at home, and from the people with whom I have shared the daily tasks of living, of whom Sue Hesington and Heather Grogan tried to cull a huge crop of commas.

Finally, I acknowledge the financial support that came from various sources, especially the Sydney Water Board, and the University of Tasmania.

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PART I

Chapters 1 - 2

INTRODUCTION

In the first chapter, Lake Burragorang is introduced, first in its historical setting, and then in relation to its current context, as one of Sydney's major water supply reservoirs, and one of comparatively few deep, man-made lakes on the Australian continent. The second chapter concerns the methods employed in monitoring the lake over a 20 year period, and then the methods specific to this essentially analytical thesis. This second chapter, in particular, should be seen as a reference text, rather than a mandatory pathway to the main body of the thesis, because it contains all the relevant methodological information, and is necessarily referred back to in subsequent chapters.

CHAPTER 1

INTRODUCTION

INTRODUCTION

The following work arose from the recognition that Water Boards around Australia systematically collect a vast amount of physical, chemical and biological data about the water bodies that they manage. In most instances the data serves an immediate purpose as far as water management is concerned but there is rarely, if ever, the time to analyse such data from any longer term perspective.

Long term studies are increasingly recognised as valuable today for their ability to provide a view of the "normal" variation of lake behaviour over time, a scale against which to view the many short term studies that have been done, and for their importance in determining the rate at which so many of the worlds lakes are being deleteriously affected by cultural eutrophication (cf Kalff and Knoechel 1978; Likens 1984). Such studies are not common in the context of world limnology and are much less common in Australian limnology. Given the present economic climate and the long lamented lack of research funds for water resources research in Australia (see Williams 1973, 1982), the establishment of projects that promise no significant return for 5, 10 or more years seems unlikely; so that the analysis of data already accumulated by water resource managers and the active commitment to maintaining continuity of such data collection offers perhaps the best alternative for research of this kind in Australia.

The data for this study, collected over a twenty year period for a deep impoundment on Australia's east coast by the Metropolitan Water, Sewerage and Drainage Board (MWS&DB) of Sydney, are analysed here mainly in terms of the physical and chemical aspects, but have the distinction of having parallel data about algae and zooplankton for the same period. This study, therefore, forms a necessary base upon which to build a detailed analysis of the biological data. An historical aside is that the early years of monitoring Lake Burragorang were directed by Dr. Hilary Jolly, who made an important contribution to Australasian Limnology (see Bayly 1976) and was the first biologist employed by a government instrumentality (the Sydney MWS&DB) specifically to contribute to water storage management (cf Bayly 1976; Williams 1973).

Aims of the Study

The aims of this study are first, to describe the range of behaviour of Lake Burragorang (on the scale of years) with emphasis on the fundamental physical and chemical stratification cycles upon which the biology of the lake is premised and second, to determine the main factors that generate this variation. A third aim is, where possible, to contribute to the efficient management of this major water supply reservoir.

Some Comments on Long Term Data

The task of taking a data set collected for the pragmatic and immediate purpose of managing a water supply and analysing it for its contribution to the understanding of long-term events presents a number of difficulties. These may be grouped under the headings "Continuity" and "Reliability".

Continuity:-

Analysis that aims at a long term perspective requires data that has been measured using a consistent technique, and for the full study period. Yet, over a twenty year period it is virtually inevitable that technological and scientific advances change the monitoring program to some extent. Both in terms of the techniques (accuracy, mechanism) for measuring certain parameters, and by the addition or replacement of parameters in the sampling program in line with changing perception of what represents an optimal sampling regime.

Examples of such changes, for the present data set, include the technological advances which made smaller and smaller mesh sizes available for phytoplankton nets (D. Cannon pers. comm.), the change in the accuracy of nitrate determinations accompanying the use of a different metallic alloy in the reduction process (in 1974; A. Robertson pers. comm.), and the addition of measurements such those for chlorophyll-a and phosphorus concentration (in 1970) to the monitoring program.

Consequently, unless a new parameter can be reliably indexed to one that has been measured for the full study period, the continuous data set consists of the minimum number of parameters determined within the study period, and is essentially constrained to the technology and perceptions of the earliest days of the study. This also applies to the number of sample sites and the depth and time frequencies of the monitoring program. For example, in this study the minimum sampling frequency of later years is imposed on the weekly samples of earlier years to maintain continuity. With this in mind, it is especially important to protect the continuity of those parameters that have been measured from the outset, either by overlapping method changes for a certain period or conducting experiments designed to provide reliable coversion factors between new and old data. There is, however, little incentive to do this within the short term goals of day to day water supply management where the extra work involved would appear to have little immediate relevance.

Reliability:-

Continuity and reliability are obviously related concerns because a patch of unreliable data represents a break in the continuity. In the present context I refer almost exclusively to human reliability in the process of data accumulation.

The changes of personnel, over time, undoubtedly introduce some bias in certain measurements either because these determinations involve a high degree of skill and experience, as in the case of algal taxonomy, or simply because subjectivity is inherent in the procedure, as in the case of Secchi disc measurements and the visual comparator system used for turbidity measurements. These factors simply affect the error that has to be attributed to certain measurements for the purpose of a long term study relative to that which may be acceptable in a shorter term investigation.

A very serious problem may arise, however, if the integrity of someone involved in the sampling program is doubtful. This situation unfortunately arose during the present study period and the highly variable data for several parameters in 1969 is attributed to this problem.

Finally, there is the perennial problem of equipment reliability. An unfortunate example, from the present study, concerns a winch with a slightly frayed cable which introduced an underestimation of about 10% in the depth attributed to chemical measurements, over a period estimated to extend from late 1976 to the end of the study (C. Heath pers. comm.). This error remains in the data set reported here but has had little effect on the interpretation of the data.

Digression

A frustrating aspect of this type of study is the restriction to a lowest common denominator in virtually all aspects. For instance, the absence of phosphorus measurements prior to 1970 while nitrates were determined for the full study period, in a lake known to derive its inflow from notoriously phosphorus poor soils, seems negligent from the present day perspective of nutrient limitation modelling, yet the commencement of phosphorus and chlorophyll measurements followed fairly rapidly on Vollenweider's (1968) early work in this field.

It would, however, be unfair not to mention that the opposite can happen. For example, the present data set includes detailed records of the outflow and lake level enabling the investigation of advective effects. The recognition of the potentially important effects of advection in water bodies is still far from universal in the literature, although nutrient budget models represent an important exception. Limnological texts contain extremely elementary accounts of any inflow/outflow effects on the thermal structure of natural lakes. There is however movement in the right direction and this must, in part, be attributed to the published studies on reservoir hydrodynamics, those effects that led to the conclusion that reservoirs "... frequently appear to violate every principle deemed proper and appropriate for bodies of standing water." (Neel 1963). The recognition that lakes and reservoirs may lie, more or less, on a continuum is coming from a better understanding of reservoir limnology and from the investigation of a more diverse sample of natural lakes. There has been and will continue to be an important feedback to classical limnology from the study of reservoirs, especially, I believe, concerning advective effects. This arises, partly, from the fact that the managers of reservoirs have always required information on the water balance of their storages so that these data are usually available, unlike studies of many natural lakes.

In the present study, I believe this represents one of the greatest strengths of the data and one that permits a quantitative assessment of the role of advection in a lake with a water retention time that is by no means measured in days, or weeks but lies, at times, within the realm of many natural lakes. In unravelling the influence of advection, a step that apparently remains to be taken is for the reservoir outflow (and inflow possibly) to be actively manipulated at the behest of scientists interested in studying these effects. In the mean time the study of advection remains below its potential, but in a system of highly variable (year to year) inflow, like Lake Burragorang, the long term monitoring of physical, chemical and biological data offers an important potential for statistical differentiation between high and low flow years.

HISTORICAL PERSPECTIVE

An assured supply of potable water has been a primary consideration in the siting of settlements in Australia from the very earliest days of colonisation. Governor Phillip's decision to move the first settlers from Botany Bay was greatly influenced by the existence of a small, spring-fed stream discharging into Sydney Cove. This stream was to be Sydney's major water supply for about the next 38 years (MWS&DB pamphlet, undated). Consecutive dry years in 1789 and 1790 induced Governor Phillip to have 3 tanks cut into the sandstone adjoining this stream to conserve some of the water from the rapidly diminishing flow. The Tank Stream derived its name from these tanks (McIllwraith 1952).

Following Governor Phillip's resignation in 1792, Sydney saw a succession of Governors, droughts and apparently ineffectual attempts to prevent the increasing pollution of the Tank Stream by various colonial effluents. One such attempt took the form of an order published in the Sydney Gazette of December 18, 1803:- "If any person whatever is detected in throwing any filth into the stream of fresh water, cleaning fish, washing, erecting pigsties near it, or taking water out of the tanks, on conviction before a magistrate their home will be taken down and forfeit £5 for each offense to the orphan fund."

Despite such orders, and the restriction of access to the stream by means of fences and military surveillance, the problem became progressively worse. In 1820, another dry year, a Gazette notice sadly concluded:- "With much pain we have lately observed individuals washing themselves in this stream of water, particularly in that part that runs centrally from King Street because that spot is almost secluded from every eye, that of curiosity excepted." (after MacIllwraith 1952). The Tank Stream had to be abandoned as a water supply in about 1826.

The Warragamba River was discovered in 1802, apparently by Ensign F. Barrallier, during an unsuccessful attempt to find a way inland across the Blue Mountains. In 1810 Governor Macquarie made a day excursion to explore this new river and directed that it be called by the Aboriginal name which he interpreted as "Warragombie". The name "Warragamba", which apparently derives from the Aboriginal words for swamp and ti-tree, seems to have been used first by Surveyor General Oxley in 1825 (after Aird 1961).

The Warragamba River was formed by the confluence of the Wollondilly and Cox Rivers just upstream of the precipitous Warragamba Gorge. Above this junction the Wollondilly flowed through the broad Burragorang Valley which eventually gave its names to the present day lake. The discovery and early occupation of this valley touches on the unlawful side of Australian history. George Boxall (1899) describes the Burragorang Valley as a natural enclosure with but a single practicable entrance, "... which, in early times, ..., was easily blocked with a few saplings, so that sheep, cattle or horses turned into the valley could not escape.". He tentatively ascribes the discovery of this entrance to a party of bushrangers seeking a passage through the mountains to a white settlement popularly thought to lie somewhere in that direction. Lying only 54 miles from Sydney, the valley afforded a secure hiding place for bushrangers over some years around the 1820s. "It was to this valley that Will Underwood and his gang were said to retire when hard pressed or when they required a rest. Underwood operated on the roads about Campelltown, Liverpool, Penrith, and Windsor, sometimes sticking-up people, and robbing farms on the Liberty Plains and other places between Parramatta and Sydney. The gang was a large one and continued to operate in the more populous districts for some two years", from 1830-32 (Boxall 1899). Underwood lasted longer than other gang members such as Johnny Donohue, Webber, and Walmsley, who were captured or shot before their leader's death in 1832.

Shortly thereafter the existence of the valley was revealed by a traitor who led the troopers there. In the ensuing fight several bushrangers were captured and the gang was broken up. Boxall (1899) explains the entry of the Burragorang Valley into Australian Folklore:- "Later on the valley came to be known, from the horrible tales told of the convicts who made use of it, as "Terrible Hollow", and under this name it is introduced by Rolf Boldrewood in his "Robbery Under Arms". Among the old hands themselves it was known as "The Camp", The Shelter", or "The Pound"..... The evil reputation which the valley had acquired, at first prevented settlement there, but when the bushrangers and their doings had been forgotten, the Government threw the valley open for selection".

In the years between 1826 and 1888, when the Metropolitan Water Sewerage & Drainage Board (MWS&DB) was first constituted to take responsibility for Sydney's water supply and sewerage, the city relied on schemes that diverted water from the Lachlan and Botany Swamps. Busby's Bore, completed in 1837 during another prolonged drought, brought water from the Lachlan Swamp to Hyde Park via a tunnel made using convict labour (MWS&DB pamphlet, undated). The water was sold for one shilling per cask (MacIllwraith 1952). By the 1850's this scheme was inadequate and in 1858 steam driven pumps began to provide water from the Botany Swamps (MWS&DB pamphlet, undated).

Attention was first drawn to the water supply potential of the Warragamba River in 1845 by Count P. E. Strzelecki. In 1869 Lieutenant T. Woore proposed the construction of a dam in the Warragamba gorge, the single narrow outlet for the extensive catchment of the Wollondilly and Cox Rivers (Aird 1961). At the time this was rejected in favour of the Upper Nepean Scheme, which was initially a project to divert water via a system of tunnels and open canals to Prospect Reservoir. As this scheme approached completion Sydney was again faced with a severe drought, and a temporary system of pipes and timber fluming was brought into operation in 1886. "Hudson's Temporary Scheme" bridged the gaps in the part-completed Upper Nepean Scheme and became redundant when the latter project was completed in 1888 (MWS&DB pamphlet, undated).

Plans to dam the Warragamba River were not seriously considered again until the early 1900's. In 1919 a dam-site was selected 3.24 km upstream of the confluence of the Warragamba and Nepean Rivers (Aird 1961). No further progress towards construction of the dam was made until the eight year drought which lasted from 1934-42. The Upper Nepean Scheme, which in 1935 included four impoundments (Cataract, Cordeux, Avon, and Nepean), proved inadequate in the face of the extended drought (MWS&DB pamphlet, undated). Consequently, an emergency project was begun in 1937 to provide water from the first impoundment of the Warragamba River. In 1940, with the completion of a 15 m weir just upstream of the earlier proposed dam-site, water was pumped to the storage reservoir at Prospect. This was the first of what was to be a major contribution to Sydney's water supply from the Warragamba catchment.

Renewed investigation of the geology of Warragamba Gorge was delayed by World-War II. However, sufficient data was eventually collected for analysis by geologists from Sydney University, who recommended a site 1.26 km upstream of the earlier chosen site. Following confirmation of the site by an eminent engineering expert from the U.S.A., the Board approved the site in October 1946 (Aird 1961). Warragamba Dam was completed in 1960. When full supply level was reached in November 1961, the total available capacity of Sydney's water reticulation system was boosted to four times that which was available prior to the exploitation of the Warragamba catchment.

CONSTRUCTION OF WARRAGAMBA DAM

The construction of the 137 m high Warragamba Dam remains the largest single engineering project performed by the Sydney MWS&DB (Flynn 1981). In the Australian context Warragamba is one of the country's largest concrete dams, and the resultant lake is among few extensive, deep bodies of fresh water on the Australian continent. Dartmouth Reservoir (Victoria) is considerably deeper (see Williams 1973), and the Middle Gordon Impoundment (Tasmania) holds a much greater volume of water (see Williams 1973). These are, however, more recent achievements. At the time of its construction Warragamba Dam represented a major undertaking for Australia, creating world-wide interest from both general and specifically technical viewpoints (Aird 1961).

Preparation of the Dam-site

Apart from small scale preliminary works, progress on the Warragamba Scheme was delayed by shortages of men, money and materials during World War II. In 1948 work began simultaneously on the new upstream coffer-dam, with an associated diversion tunnel blasted through the eastern bank, and on the task of excavating a deep trench across the gorge (Flynn 1981). The preparation of a foundation area for the dam took 5 years and necessitated the removal of unsuitable material followed by a considerable amount of drilling and grouting of seams and joints in the bedrock. A fault zone underlying the western side of the valley required special attention, and much of the faulted rock was replaced with concrete (Aird 1961).

Flooding of the construction site

The upstream coffer-dam and the diversion tunnel were designed to cope with normal river flow and minor floods (up to c. 8.8 x 10^6 m^3 day^{-1}; 102 m^3

sec^{-1}). The decision to accept occasional flooding of the work-site was made because of the cost of diverting the sometimes massive outpourings of the 9000 km^2 catchment. Consequently, it was imperative to provide early warnings of impending floods, particularly as it was possible for them to occur when there had been no rainfall at the construction site (Aird 1961). The task was made all the more difficult by the irregularity of the rainfall and its lack of pronounced seasonal pattern. About six hours was required to remove the workmen and equipment from the valley, in advance of a flood. The responsibility for this early warning fell to the continuously - manned - river - control - office, where reports of the levels of the Wollondilly and Cox Rivers and rainfall information from various parts of the catchment were co-ordinated to establish the flood risk. Arrangements were made with postmasters, the manager of the Jenolan Caves Hotel and others to phone the office at Warragamba if a fall of 1 inch or more occurred in their locality (Aird 1961). This network stretched from the source of the Wollondilly (between Taralga and Crookwell), to Wallerawang in the Blue Mountains where the Cox River rises. A storm in the Taralga district could take up to 27 hours to meander the 418 km length of the Wollondilly and reach the dam. In contrast, a fall in Wallerawang district would travel only 185 km in the Cox River valley (Aird 1961).

Flooding of the dam-site was common, particularly in the early years. Thirty nine floods occurred between 1952 and 1956. In the wake of the larger of these floods, as much as 75000 m^3 of sand and silt had to be removed from the site (Flynn 1981).

Building the Dam

Following completion of the coffer-dam the works area was ready to be pumped dry by 1951. Around this time the task of shifting aggregate for the concrete was begun. This material was moved from McCanns Island (4 km downstream of Penrith on the Nepean River) by means of an aerial cableway which transported up to 175 tonnes hr^{-1} to Warragamba, 20 km away. The first concrete of the wall was placed in 1953 (Flynn 1981). To minimise cracking of the cement, the dam was constructed as separate blocks which were later grouted together. Prior to this final sealing process the concrete had to be cooled to the estimated final stabilised temperature. As a first step, ice was added to the concrete mix to limit the rise in temperature inherent in the chemistry of the setting process, the first use of this technique in Australia. To further speed the temperature stabilisation, coils of mild steel tubing were embedded in the concrete and chilled water was circulated for several months after the initial pouring. Combined, these techniques shortened the stabilisation period from c. 100 years (Aird 1961) to a few months. Shrinkage joints between the blocks were subsequently filled with fine cement grout pumped in under pressure. This made the wall act as a continuous monolithic block (Aird 1961). Work on the dam wall continued 24 hours a day, 7 days a week. As the wall increased in height, it became possible to minimise the overall interuptions due to flooding. A gap in the spillway section permitted most floods to pass harmlessly, while work continued at the upper levels (Flynn 1981). At the lower levels work was still disrupted periodically by floods, particularly the construction of the dam-apron which commenced in 1956. The apron and downstream training walls dissipate the energy of flood flows and protect the dam from undercutting.

"In 1957 the dam wall was high enough to hold water for delivery to Prospect Reservoir, via pipeline, and the diversion tunnel was closed off" (Flynn 1981). This represents the second major step in the formation of Lake Burragorang. From 1957 the level, and therefore extent of the lake, was controlled by the height of the dam wall and not by the relatively diminutive coffer-dam, as had been the case since 1940. The lake may, in fact, have been independent of the height of the coffer dam earlier than this. An MWS&DB

employee who joined the workforce at Warragamba in 1955 suggested to me that the coffer dam was not controlling the lake level even then (W. Mayrhoffer pers. comm. 1982), and Jolly (1966a) comments that water storage in Lake Burragorang started in 1956.

Work on the dam was temporarily disrupted in 1957 by a major fire which burnt a considerable area of catchment and threatened Warragamba Township. Flynn (1981) observes, "Predictably, a flood occurred soon afterwards. A foul black rush of water from the fire ravaged catchment brought with it the remains (and the odour) of burnt livestock and native wildlife." In 1958 construction of the dam proceeded swiftly, with a record 2080 m^3 of concrete being poured in a single 24 hour period (Flynn 1981). Warragamba Dam was officially opened on the 14th of October 1960. Lake Burragorang was slower to form, but on February 12th 1959 the water reached 79.9 m above sea level (36.8 m below Full Supply Level; FSL), giving a maximum depth of 68 m adjacent to the dam wall, and gravitational delivery to Prospect Reservoir commenced at a rate of just over 0.364 10^6 m^3 day^{-1} (Aird 1961). The lake was finally filled by the major flood of November 1961 which raised the water level the last 4.49 m to flood over the dam crest, and is the single largest flood in the present study period (1961 - 1980). Fig. 1.1 shows the approximate extent of the lake at some important stages of its development, 1940, 1959 and at FSL.

Preparation of the Lake Basin Prior to Flooding

Concurrent with the construction of Warragamba Dam, there were a variety of measures taken to prepare the Cox and Wollondilly River valleys for flooding.

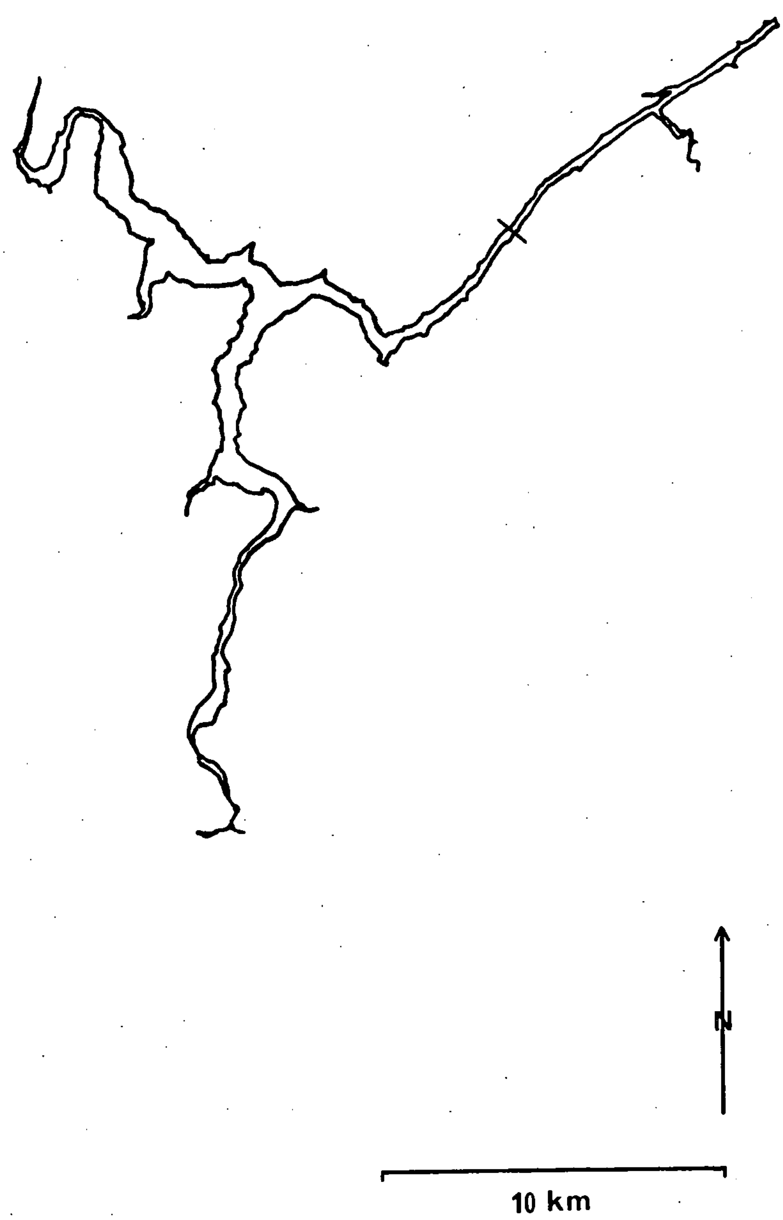
All properties in the area to be inundated, and those within 3.24 km of the proposed lake, were purchased by the MWS&DB. The Burragorang Valley, in particular, had a well established farming community. "The Burragorang

FIGURE 1.1

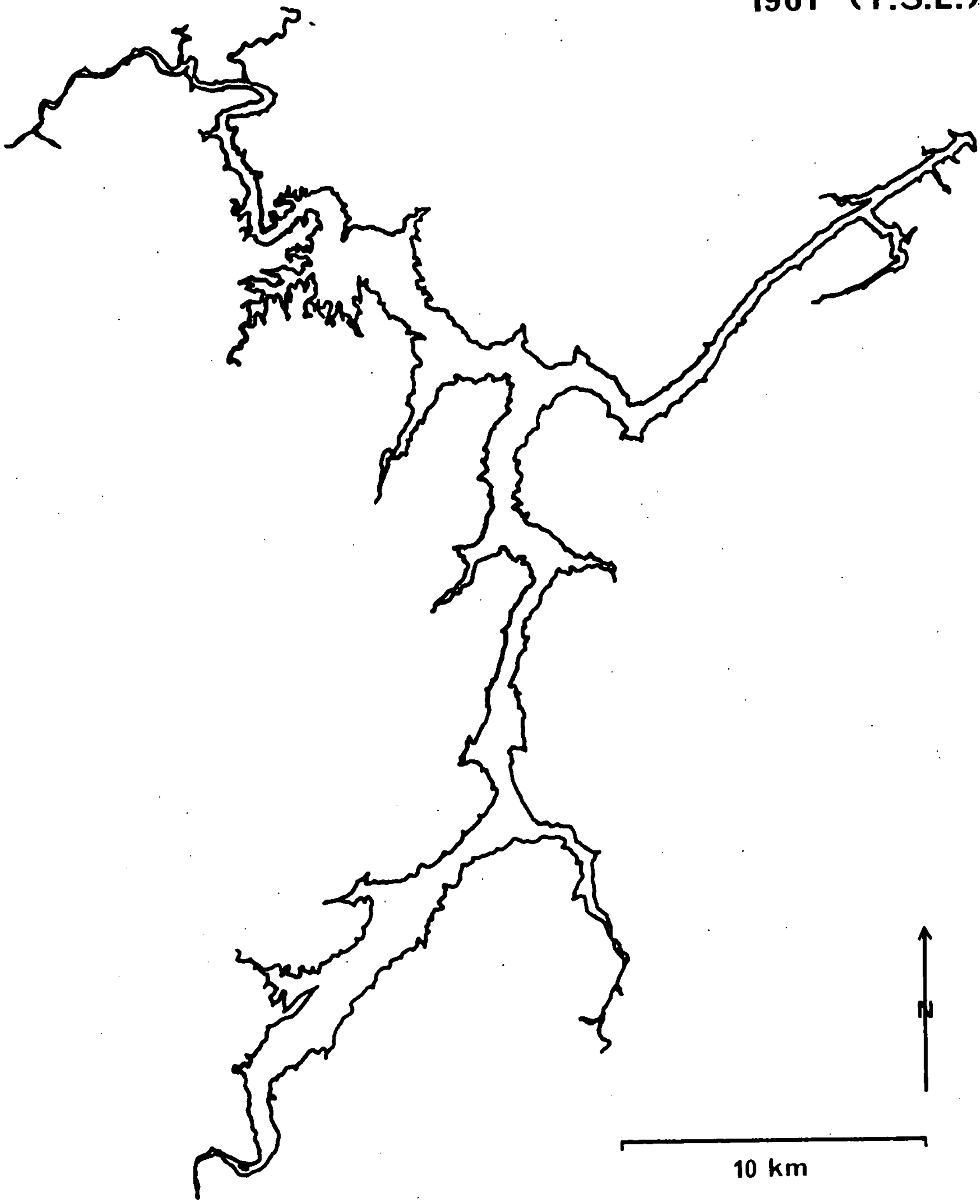
The following maps, show the extent of Lake Burragorang, at some significant stages in its development. The left hand map shows the lake as it was in February 1959, when a relative level of 79.9 m was first reached, and gravitational delivery of water to Prospect Reservoir, began. A line marks the furthest extent of the lake formed behind a 15 m high weir constructed near the present dam-site in 1940. The right hand map shows the lake at full supply level, which was first reached in the floods of November 1961.

Note: The two maps were derived from different maps, of varying scale (the 1959 outline comes from a pre-impoundment map done in the 1950s, at a scale of 1 chain to the inch), and in the serial reductions (photocopied) necessary to produce the final maps, some slight distortion of the outline has occurred, making a direct overlay impossible.

1959



1961 (F.S.L.)



Valley and the valley of the Wollondilly River further south were some of the earliest settled areas of N.S.W., both due to their proximity to Sydney and location on one of the early routes south of the colony." (Soil Conservation Survey 1965). This population and a lesser number from the Cox River valley were relocated and all the abandoned buildings were demolished (Aird 1961).

The area to be flooded was systematically cleared of trees and understorey vegetation. This involved c. 500 men, of which 200 were Water Board employees and the remainder were private contractors. The work began in the early 1950's and continued until 1959. Trees were either pushed down and burnt, or cut at about half a metre above the ground and the stumps painted with 245T (in oil) to retard regrowth (M. Whooten pers. comm. 1982). Some timber was taken for use at Warragamba Township and in the construction of the dam. The remaining timber and cleared undergrowth was burnt in situ. Water Board employees were used to clear some of the more difficult terrain, particularly in the Warragamba Gorge where steep slopes and dense wet-schlerophyll forest (in shaded gullies) made the task an arduous one (Whooten pers. comm. 1982). In the time between clearing and flooding, some regrowth inevitably occurred. Whooten (pers. comm. 1982) estimated that this included seedling trees up to about 2 m in height, prior to final inundation.

OPERATION OF WARRAGAMBA DAM

Warragamba is a "Straight gravity-type" dam, which discharges floodwater directly over the dam crest (Aird 1961). The crest is 350.5 m across and has a 94.5 m wide spillway at its centre. The spillway is partitioned into five openings which contain the four radial gates and the centrally placed drum-gate which has a 27.4 m clear span. This gate has the form of a long, hollow steel drum hinged along its upstream lower edge to the upstream crest (Aird 1961). The drum floats in a large chamber and is operated as the chamber is filled or emptied of water. Removing water from the drum-gate chamber causes the gate to subside into the chamber, forming a continuation of the crest profile when fully open; water from the lake then flows over the top. The radial gates (each 12.2 m clear span and 13.3 m deep) are supported on pivoting radial arms and open by lifting upwards, allowing water to pass underneath (Aird 1961).

There are four 2.13 m diameter outlets from the dam. A low level offtake (93.4 m below FSL), which supplied water to the pumping station until 1959, and three situated at 61.6 m below FSL which now receive water via movable screens, permitting selective offtake from the lake surface down to the level of these outlets (ie continuous selection of withdrawal level from 0 m - c. 60 m below FSL). This variable offtake is used to supply water to Prospect Reservoir and eventually to Sydney. A fixed level offtake, centred at 44.5 m below FSL, supplies water to the hydro-electric power station. This outlet consists of three vertical slits, covered by screens, in a semi-circular array protruding about 3 m - 4 m from the dam wall. Each screen is 4.3 m wide and 18.3 m tall, so that the offtake structure extends vertically from 35.4 m below FSL to 53.6 m below FSL (L Petri pers. comm. 1984). The hydro-electric offtake has the greatest volume capacity of the subsurface outlets and is capable of withdrawing about 5.5 10^6 m^3 day^{-1}, compared to a maximum capacity of c. 2.5 10^6 m^3 day^{-1} for the variable, water supply offtake.

LAKE BURRAGORANG AND ITS CATCHMENT

The Lake

Table 1.1 contains information on the morphometry of Lake Burragorang. The lake is narrow, elongate and dendritic in form. It has a very high development of shoreline, relative to many natural lakes, indicating a considerable potential for the effect of littoral processes in the lake

Table 1.1 Selected data for Lake Burragorang.

Position of lake centre		33° 55' S., 150° 25' E.
Lake surface area (A_0)		75 km^2
Mean width		1.44 km
Catchment area (total)		9013 km^2
Catchment/Surface area ratio		120.25
Shore line (L)		354 km
Development of shore line ($L/2\sqrt{\pi A_0}$)		11.53
Surface elevation		116.7 m A.M.S.L. at full supply level
Lake volume (when full)		2057 x 10^6 m^3 before siltation
Volume development (z/z_{max})		0.27
Full supply level first achieved		November 1961
Maximum depth (before siltation; z_{max})		105 m
Mean depth (z)		27.4 m
Relative depth ($50\ z_{max} \sqrt{\pi}/\sqrt{A_0}$)		1.07
Retention time (lake volume/inflow-evaporation), 1962 - 1980		
	range	0.5 - 22.8 years
	mean	4.8 years
	median	1.8 years

(Hutchinson 1957). This figure (11.53) is, however, comparable with other impoundments, such as Lake Brokopondo and the large African impoundments (Lakes Kariba, Volta, Nasser-Nubia, Kainji, Cabora Bassa, and Kossou; see Heide 1982). In comparison with natural lakes, Lake Burragorang has a very small volume development (mean depth/maximum depth; Hutchinson 1957), but is within the range of values reported for other impoundments Heide (1982). The relative depth lies within the fairly broad range reported by Hutchinson (1957) for natural lakes, being intermediate between the Fjord type lakes (high relative depth) and most of the tectonic and glacial lakes (lower relative depths). Interestingly, the impoundments reported by Heide (1982) all have much smaller relative depths than Lake Burragorang. A feature of some importance is the large catchment area to surface area ratio which clearly indicates that events in the catchment will exert a profound effect on the lake. The annual retention time for the lake is quite variable, in keeping with the variability of stream flow typical of Australian conditions (see Williams 1982).

Table 1.2 gives data on the distribution of volume with depth in the lake. This clearly shows the small contribution of the deeper layers (> c. 30 m) to the total lake volume. The lake area at 1 m intervals (estimated by curvilinear regression, below 45 m), is given in Appendix 1.

The Catchment

In Fig. 1.2 Lake Burragorang is shown in relation to Sydney and Australia's east coast. The oceanic influence on the lakes thermal pattern will be discussed in a later section. Fig. 1.2 also shows the total catchment drained by the Wollondilly and Nattai Rivers in the south, and the Cox and Kowmung Rivers in the north and west. The flow in these four rivers is gauged, and accounts for about 83% of the total inflow (D. May pers. comm. 1983). The catchment of the Wollondilly River (c. 4800 km^2; May pers. comm.) comprises

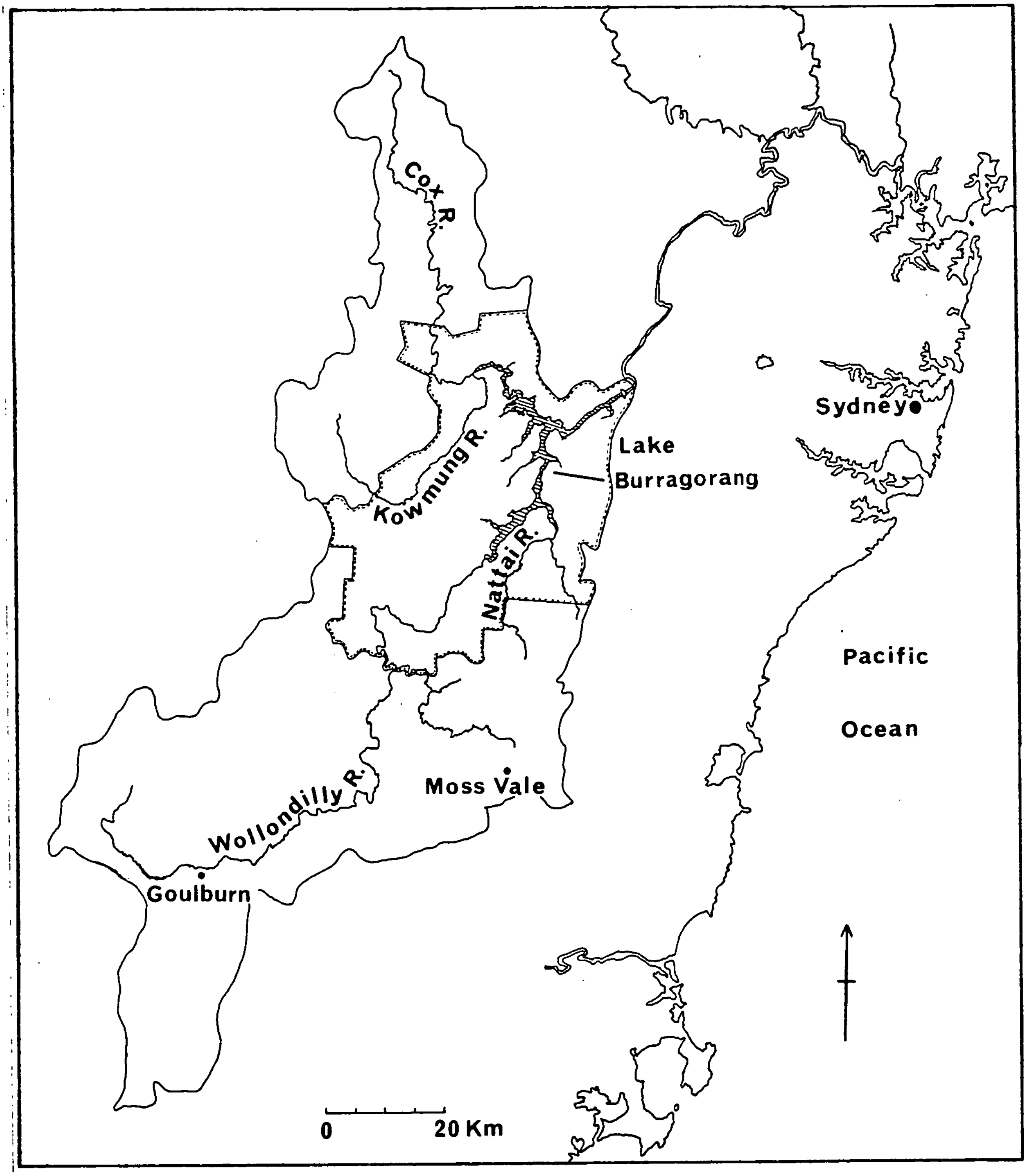
Table 1.2 Volume distribution with depth in Lake Burragorang.

Depth range	Volume x 10^6 m^3	% of Total
0 - 6 m	420.248	20.4
6 - 12 m	362.958	17.6
12 - 18 m	308.169	15.0
18 - 24 m	256.993	12.5
24 - 30 m	210.766	10.2
30 - 36 m	169.144	8.2
36 - 42 m	129.588	6.3
42 - 105 m	199.134	9.7
Total:	2057.000	

FIGURE 1.2

A map of present day Lake Burragorang (at FSL), showing the physical catchment, and the "Inner catchment" (dotted line) which comes under the jurisdiction of the Metropolitan Water Sewerage and Drainage Board of Sydney (MWS&DB). The Public are allowed limited access to the inner catchment, for instance for bush-walking, but are not allowed inside a 3 Km wide area surrounding the lake, or onto the lake itself.

Two major river systems drain into the lake, the Cox River - Kowmung River system drains the area to the west of the lake, while the Wollondilly River - Nattai River system, drains the region generally south of the lake. May (pers. comm. 1983) estimates that these four rivers account for about 83% of the total inflow into Lake Burragorang.


Cox R.
Kowmung R.
Lake
Burragorang
Nattai R.
Sydney
Pacific
Ocean
Wollondilly R.
Moss Vale
Goulburn
0
20 Km

53% of the total physical catchment area. The Wollondilly and Cox Rivers are chemically distinct, in keeping with the different geology and topography of their respective catchments. The inner catchment, that area under the direct statutory control of the MWS&DB, is also shown.

Table 1.3 contains some basic information about the catchment including its land-use (after Bowen and Smalls 1980) and fire history since the late 1950's. A relatively high percentage of the catchment has been retained with the natural dry sclerophyll vegetation and the lake is surrounded by 3000 km^2 of land (inner catchment) in which public access is restricted. The public are not allowed within 3 km of the lake itself; a statute that is reinforced by the precipitous terrain around much of the lake shoreline. No recreational use of the lake is permitted.

Topography:-

The following impression draws on the Soil Conservation Sevice reports (1962, 1965), the Sydney and Wollongong 1:250000 Geological Series Sheets (SI 56-5 and SI 56-9; 1966) and explanatory notes compiled by Bryan et al (1966), and the 1:100000 World Aeronautical Chart (3456, Sydney; 1978).

The catchment is bordered on the west by the Great Dividing Range which rises to about 1300 m at the western extremity of the catchment. The lake itself is deeply entrenched in the Blue Mountains Plateau which is bounded on the west by the Cox River; Warragamba Dam is close to its eastern boundary, where the plateau falls sharply to the Cumberland Plain. The plateau is about 180 m high at the dam but much of the lake is surrounded by peaks of 600 - 700 m in height.

The Cox and Kowmung rivers have cut from 600 - 900 m down into the plateau, forming the impressive cliffs characteristic of their dissected catchment system north and west of the lake. South of the lake the Wollondilly River catchment is mainly described as hilly and undulating, but there are places where the river has cut deeply into the plateau forming

Table 1.3 Catchment information; land-use and fire history.

Catchment area	total	9013 km^2	
	inner	3000 km^2	
Land use in catchment (total)			
Native dry sclerophyll forest		50%	
Native pasture		30%	
Improved pasture		19%	
Urbanized		<1%	
Catchment fire history: (Fire years June - June). Time of major fires, and approximate area of the catchment that was burnt.			
1957-58	(Sept.- Dec.)	1000 Km^2	(Inner catchment)
1964-65	(March)	1800 Km^2	(Inner catchment)
1968-69	(Oct.- Nov.)[a]	1400 Km^2	(Inner catchment)
1978-79		1200 Km^2	(Outer catchment)

[a] Probably a greater area of the total catchment was burnt in 1968 than in 1965 (M. Whooten pers. comm. 1982)

precipitous gorges. The plateau surface rises from about 500 m near the Nattai River to about 900 m in the southernmost extension of the catchment.

Geology:-

The following information is taken from the Sydney and Wollongong 1:250000 Geological Series Sheets (SI 56-5 and SI 56-9; 1966) and explanatory notes compiled by Bryan et al (1966). A simplified geological map of the region surrounding lake Burragorang is given in Fig. 1.3. The map does not include the whole catchment, particularly that of the Wollondilly River, but covers an area somewhat greater than the inner catchment.

The lake occupies a region in which the Wollondilly, Cox, and Warragamba Rivers have cut deeply into the Triassic and Permian sedimentary rocks, comprising four layers. The capping of Hawkesbury Sandstone overlies sandstone and shale of the Narrabeen Group, below which are the Permian sediments of the Illawarra Coal Measures, and then a comparat~~ely~~ively thick layer of siltstone, shale and sandstone of the Shoalhaven Group (Berry Formation). The Cox River arm of the lake, touches on an area of Upper Devonian quartzite and sandstone of the Lambie Group.

Rocks of the Lambie Group and Lower Carboniferous granitic rocks cover much of the Cox - Kowmung River catchment area, with a small region of Silurian slate, phyllite and sandstone near the southern extreme of the Kowmung River drainage basin.

The Nattai River catchment is similar to the area occupied by the main body of the lake, but the Wollondilly River drains a more geologically diverse area and one that differs from much of the rest of Lake Burragorang's catchment. Within the bounds of Fig. 1.3 the Wollondilly River crosses three regions. Closest to the lake, is an area of Upper Devonian Bindook Porphyry; south of this is a region of Devonian granite, and a small area of Ordivician slate, quartzite and phyllite.

<u>FIGURE 1.3</u>

A geological map covering Lake Burragorang (in black) and much of its catchment. Simplified from the SYDNEY (1:250,000) Geological Series Sheet SI 56-5 and WOLLONGONG (1:250,000) Geological Series Sheet SI 56-9.

Key:-

A. Triassic: Mainly Hawkesbury Sandstone, with a fringe of Narrabeen Group sandstone, which occupies a larger region north of the lake, and an area of Wianamatta Group shale in a band down the eastern margin of the map, and (with outcropping Tertiary basalt) in the region south of the lake.

B. Permian: Mainly siltstone, shale and sandstone of the Shoalhaven Group (Berry Formation), fringed by rocks of the Illawarra Coal Measures (which occupy a larger region south of the lake).

C. Lower Carboniferous: Granitic rocks (adamellite, granite, and granodiorite) in the region drained by the Cox River, and with outcroppings in the D region on the map.

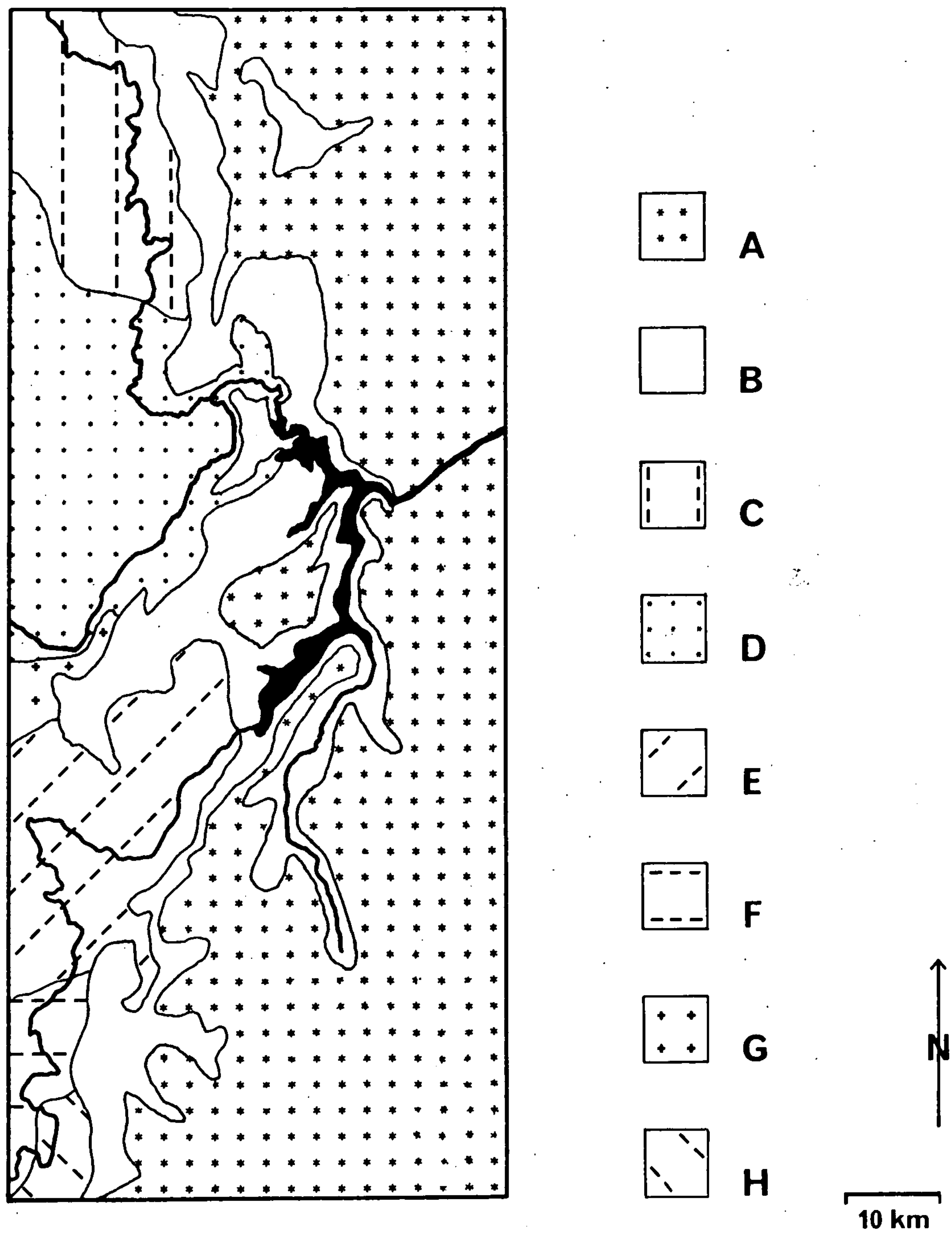
D. Devonian: Mainly Upper Devonian quartzites and sandstones of the Lambie Group.

E. Upper Devonian: Bindook Porphyry (porphyry, dacite and tuff).

F. Devonian: Granite, granodiorite and porphyry, with an outcrop of Megalong Conglomerate (Permian) in the south.

G. Silurian: A small region of slate, phyllite, sandstone and limestone, lying south of the Kowmung River.

H. Ordivician: A small region of slate, quartzite and phyllite, drained by the Wollondilly River.


A
B
C
D
E
F
G
H
N
10 km

Soils:-

This information comes mainly from the Soil Conservation Survey (1965), and is indexed to the key used for Fig. 1.3. The development of soils is conditioned by the topography in many areas of the catchment.

A. Both the Hawkesbury and Narrabeen Group sandstones generally produce sandy skeletal soils and a variety of sandy podsolic soils of low fertility, although Beadle (1962) distinguished between the two groups of parent rocks, on the basis of phosphate content, describing the Hawkesbury Sandstone as "... by far the lowest in phosphate.". The Wianamatta Group shales produce fairly deep soils a little more fertile than those derived from the sandstone. The small areas of outcropping Tertiary basalt form deep, fertile chocolate and reddish chocolate soils.

B. These Permian sediments have produced soils similar to those derived from the Triassic sediments (see A.), although the range of podsolic soils derived from the rocks of the Illawarra Coal Measure contains some of moderate fertility.

C. The Lower Carboniferous granites give rise to coarse textured yellow podsolic soils, which are susceptible to gully erosion.

D. The Upper Devonian quartzites and sandstones of the Lambie Group, form sandy skeletal soils with pockets of podsolic soil where topography permits.

E. The Upper Devonian Bindook porphyry forms heavy textured podsolics (grey and red) of moderate fertility. This region contains the most severely eroded soils of the catchment.

F. The Devonian granites have formed soils similar to those of the Lower Carboniferous granites (see C.), coarse podsolic soil of low fertility.

G. These Silurian slates and sandstones produce red podsolic soils of moderate fertility where the topography permits.

H. The Ordivician slates, quartzite and phyllite form fine textured podsolic soils of moderate fertility.

The predominance of relatively poor soil in the catchment of this lake, and the very low phosphate content of the soils derived from the Hawkesbury sandstone (Beadle 1954, 1962), is likely to be significant in relation to the nutrient content of the water run-off from the catchment of the lake.

Some general observations from the Soil Conservation Service reports (1962, 1965) are of interest in relation to sedimentation in the reservoir. The Wollondilly River is regarded as carrying a considerable load of fine clay sediment during floods, in direct contrast to the Cox River. There are, however, some reasons to suppose that the erosion, particularly in the inner catchment, will have decreased in the period since the MWS&DB took charge. First, because grazing sheep have been excluded from the area altogether as their intensive grazing habits threaten ground cover and grazing generally has been restricted. Second, because of improved fire control throughout the region and third, as a result of improvement in agricultural methods for soil conservation, in part assisted by the influence of "city capital". Interestingly, in the area of the Wollondilly catchment immediately south-west of the lake's southern end a considerable area that was used for grazing has reverted to eucalypt forest, after the depression (1930's) and the increasing rabbit population drove many farmers from the area.

Vegetation:-

This information comes from the Soil Conservation Service reports (1962, 1965) and I have not checked the nomenclature for changes in usage. Their information appears to have been substantially derived from Pidgeon (1937, 1938, 1940, 1941). A More recent publication, of direct relevance, is that of Black (1982) who gives a detailed account of the area close to the lake (north, west, around to south west) including much of the Kowmung River catchment and the innermost portion of the Cox River catchment. Beadle (1954, 1962) gives a general account of vegetation distribution in the Sydney district. I have not drawn on these publications.

In the region surrounding the lake there are a variety of dry sclerophyll eucalpyt associations, including the following species:-

Eucalpytus gummifera	(Red bloodwood)
E. piperita	(Sydney peppermint)
E. eugenioides	(Stringybark)
E. racemosa	(Snappy gum)
E. punctata	(Grey gum)
E. sieberi	(Black ash)
E. blaxlandi	(Mountain stringybark)
E. macrorhyncha	(Red stringybark)
E. rossii	(Snappy gum)
E. erebra	(Ironbark)
E. hemiphloia	(Grey box)
Angophora costata	(Smooth-barked Apple)

on the sandstone plateau surface and:-

E. eugenioides	
E. punctata	
E. tereticornis	(Forest red gum)
E. numerosa	(Peppermint)
Angophora floribunda	(Rough-barked apple)
Casuarina cunninghamiana	(River she-oak)
Acacia decurrens	(Black wattle)

on the slopes of the underlying Permian sediments and on the river banks.

In protected moist gullies, wet sclerophyll eucalpyt forest occurs which includes:-

E. fastigata	(Brown barrel)
E. viminalis	(Ribbon gum)
E. dalrympleana	(Mountain gum)

E. pilularis	(Blackbutt)
Syncarpia sp.	(Turpentine)
Melaleuca sp.	(Paperbark)

and some of these species are also found in the wetter upland regions of the Cox and Kowmung River catchment, most of which remains under native forest as a result of its natural inaccessability and poor soils. There are some patches of sub-tropical rainforest in protected deep gullies.

In contrast, much of the Wollondilly River catchment (apart from the inner catchment) has been under various forms of agriculture for many years and little of the native forest remains. There are areas of savannah woodland which include:-

E. pauciflora	(White sally)
E. stellulata	(Black sally)
E. melliodora	(Yellow box)
E. blakleyi	(Blakley's gum)

Where the land has been cleared, the resultant grassland includes species of Poa, Themeda, Danthonia and Stipa, while introduced pasture grasses include Ryegrass, Cocksfoot, White Clover and Subterranean Clover.

CLIMATE

Again, with the exception of inflow, evaporation and solar radiation figures, much of the following information comes from the Soil Conservation Service reports (1962, 1965). In the broadest terms, the climate may be regarded as warm temperate, varying to cool temperate at higher altitude.

Rainfall and Inflow

Rainfall varies considerably within the c. 9000 km^2 catchment, the long term mean ranges from about 500 mm yr^{-1} to about 1600 mm yr^{-1} in various regions that drain into the lake. The long term mean rainfall for the catchment is about 820 mm yr^{-1} with an annual catchment yield of about 120 mm yr^{-1} (c. 15%; Soil Conservation Service 1962, 1965). The rainfall is highly variable, both within and between years and there is no pronounced seasonality. These points are demonstrated by the monthly records of total inflow minus evaporation, presented in a later section, and the fact that the annual water retention time for Lake Burragorang varies between 0.5 and > 22 years (Table 1.1) and has a coefficient of variation of 139%. The mean evaporation is about 1140 mm yr^{-1} (measured over 7 years).

It is instructive, however, to examine the frequency distribution of monthly inflow totals (minus evaporation) for the period from August 1961 to December 1980. There have been 3 inflows greater than 1000 10^6 m^3, in March, June, and November. The total of 7 inflows > 750 10^6 m^3 are distributed in two groups, March - June (a total of 5) and November (2). Including inflows > 500 10^6 m^3 extends the March - June group and adds in January (2), leaving the November frequency unchanged. For inflows > 250 10^6 m^3 all months of the year are represented, but the months from July - October appear to be under-represented.

It seems, therefore, that for the larger inflows (> 500 10^6 m^3) there may be a slight autumn - early winter (March - June) dominance, followed by a mid winter - early spring period (July - October) when they are less frequent.

The mean monthly evaporation, for the period 1962 - 1980, ranged from 3.1 10^6 m^3 (August) to a maximum of 7.9 10^6 m^3 (December) and consequently could not significantly bias any of the above inflow (minus evaporation) classes.

Solar Radiation

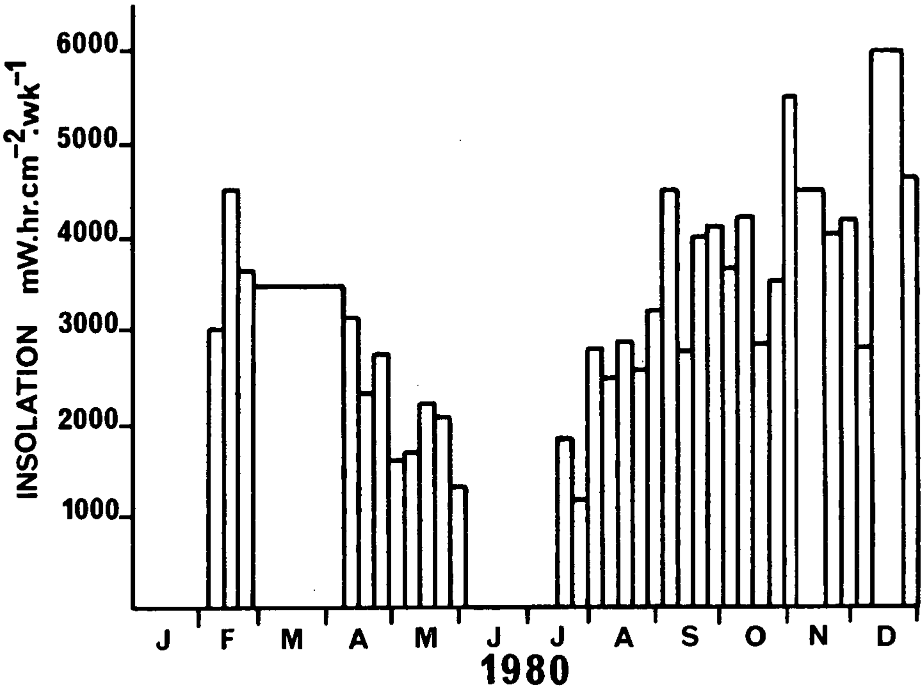
Fig. 1.4 shows records of solar radiation (400 - 1100 nm) for 1980 and 1981, measured using a RIMCO-C.S.I.RO. (R/SOLA) integrating pyranometer set

FIGURE 1.4

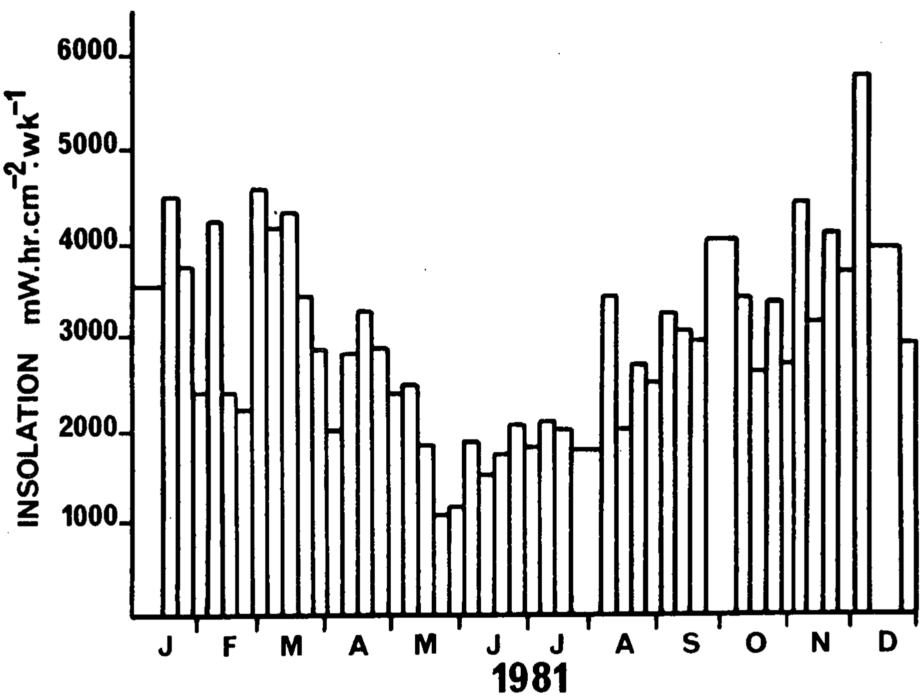
Solar irradiance (400 - 1100 nm) from February 1980 to December 1981, measured with an integrating pyranometer set up at the Warragamba Chemistry Laboratory (near to site 3D). The bars represent the total input of radiation (mW hrs cm^{-2}), corrected to a weekly value. In some cases this is averaged over several weeks, because the meter was not always read on a weekly basis. The meter was placed in as open an area as could be found, close to the Chemistry Laboratory, but some nearby Eucalypt trees cast shadows over the sensor in the late afternoon.

Note:-

> In 1980 there are two gaps in the record. The pyranometer was not set up until early February, so that data for January and the first week of February are missing. The second gap is in June and early July.



INSOLATION mW.hr.cm^-2.wk^-1
6000
5000
4000
3000
2000
1000
J F M A M J J A S O N D
1980



INSOLATION mW.hr.cm^-2.wk^-1
6000
5000
4000
3000
2000
1000
J F M A M J J A S O N D
1981

up at the Warragamba Chemical Laboratory. Each bar represents the total radiation accumulated in the appropriate period. For these two dry years the peak of solar radiation input occurred in December and the minimum in June or July.

Temperature

Mean annual temperatures range from about 11.1°C to 16.7°C, depending mainly on altitude. Minor snowfalls occur in the Cox - Kowmung River catchment area, and frosts are possible in up to 11 months of the year in places. In the Wollondilly catchment conditions range from areas in which mean daily temperatures never fall below 6°C (regarded as limiting pasture production; Soil Conservation Service 1965), to areas in which air temperatures < 0°C (severe frosts) are prevalent (usually May - September).

Unfortunately, I have no records of wind speed at or near the lake.

CHAPTER 2

MATERIALS AND METHODS

INTRODUCTION

All the chemical, analytical and computational methods for the thesis are grouped within this chapter.

Methods used in the monitoring of Lake Burragorang are presented first and then the methods used in the present analysis of the data. In this latter section some commentary is included, where this is considered relevant. Where possible, the headings within each of these sections are ordered as they will be encountered in the subsequent chapters.

Finally, there are two sections that concern individual chapters. The first, about Chapter 5, details the analytical experiment conducted to compare wet and dry years from the study period. The second, about Chapter 7, is taken with only minor changes from the paper (Ferris and Tyler 1985), which forms the basis for the section about total phosphorus-chlorophyll relationships in Lake Burragorang. There is a minor amount of repetition as a result.

DATA COLLECTION

Sampling Program

Personnel of the Metropolitan Water Sewerage and Drainage Board (MWS&DB) of Sydney have collected physical, chemical and biological data for Lake Burragorang since the late 1950's. The present study, which rests on a relatively consistent sampling program having its first full year of operation in 1961, covers a twenty year block (1961 - 1980) of mainly physical and chemical measurements.

Both the frequency of sampling and number of sample sites have changed since 1961, with a recent trend towards less frequent sampling and at fewer sample sites. In earlier years up to 5 stations in the Wollondilly arm, 5 in the

Cox arm and 2 in the narrow Warragamba gorge were sampled at weekly intervals, but regular, detailed sampling has been maintained for the full twenty year study period, at only 3 sites. These are in the Warragamba Gorge (site 3D, adjacent to Warragamba Dam) and the Cox River arm (Kedumba Creek, and Cox River above Kedumba Creek) of the reservoir (Fig. 2.1). Data for these sites include biological as well as physical and chemical parameters. The present study is almost exclusively concerned with the data for site 3D, as it is the site of most direct relevance to water quality management and has the most complete data set. The two Cox River sample sites are included in an analysis of the relationship between chlorophyll-a and total phosphorus, in Chapter 7. Fig. 2.2 is a schematic representation of the lake showing distances (along the old river courses) between sample sites, to assist in visualising the lake, particularly in relation to the distance travelled by inflows penetrating to site 3D.

In the period 1961 - 1979 data was collected from site 3D on 40 - 52 occasions each year, while in 1980 this was reduced to 26 sampling dates yielding an interval of about 2 weeks between samples. This applies to profiles of temperature, dissolved oxygen, turbidity and chloride concentration, except for 1972 - 1974 when chloride measurements were irregular (averaging about 25 samples yr^{-1}). In general, the profiles have a depth resolution of c. 6 m in the upper 36 m of the water column and c. 12 m below that depth. This gives a basic set of 11 sample depths (relative to the lake surface) of 0, 6, 12, 18, 24, 30, 36, 48, 60, 72 and 82 m approximately (following metric conversion in 1975). A variety of intermediate depths are occasionally represented in the data files prior to 1970, but the 11 standard depths form the basis of the study. The deepest sampled depth has varied considerably, from 48.8 m to 91.4 m relative to the lakes surface. The standard, 82 m, lies about 20 m above the greatest depth of water in the lake. Unfortunately, during periods of flood when boats are not allowed at site 3D

FIGURE 2.1

A map of Lake Burragorang (at Full Supply Level), showing sample sites. Distances between these sites, along the old river course, are shown in Fig. 2.2. The present study concerns site 3D almost exclusively, with some consideration of regularly sampled sites in the Cox River arm (Kedumba Creek and Cox River above Kedumba Creek). However, the inflow sites, in both the Wollondilly and Cox Rivers are those at which the river temperatures are measured, and these are compared in the text with the surface and bottom temperatures at site 3D to predict the occurrence of interflow and underflow at Warragamba Dam. The total distances traversed by inflows, following the drowned river courses, between the Cox and Wollondilly River inflow sites and site 3D, are 52 km and 57 km respectively (see Fig. 2.2).

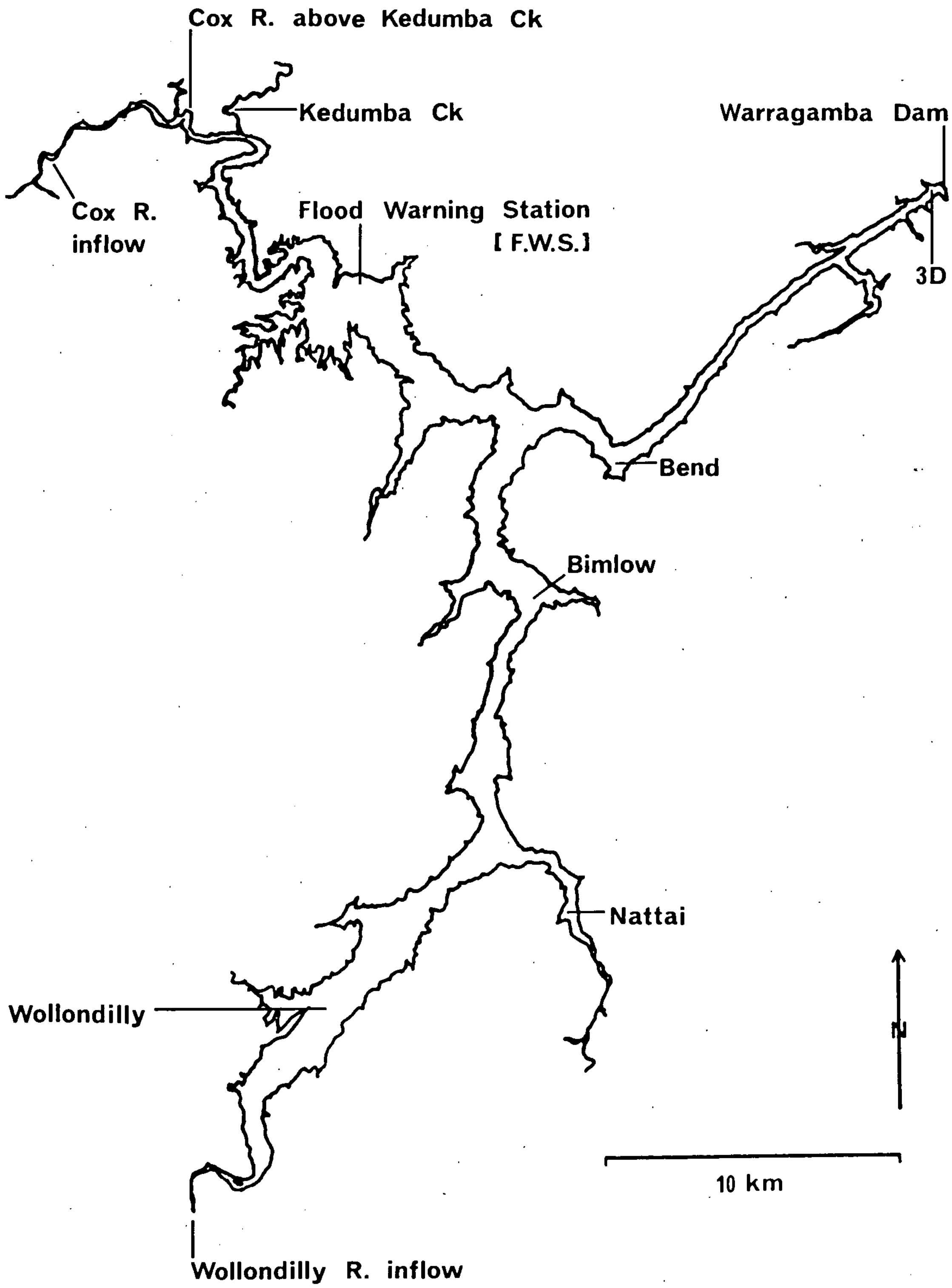
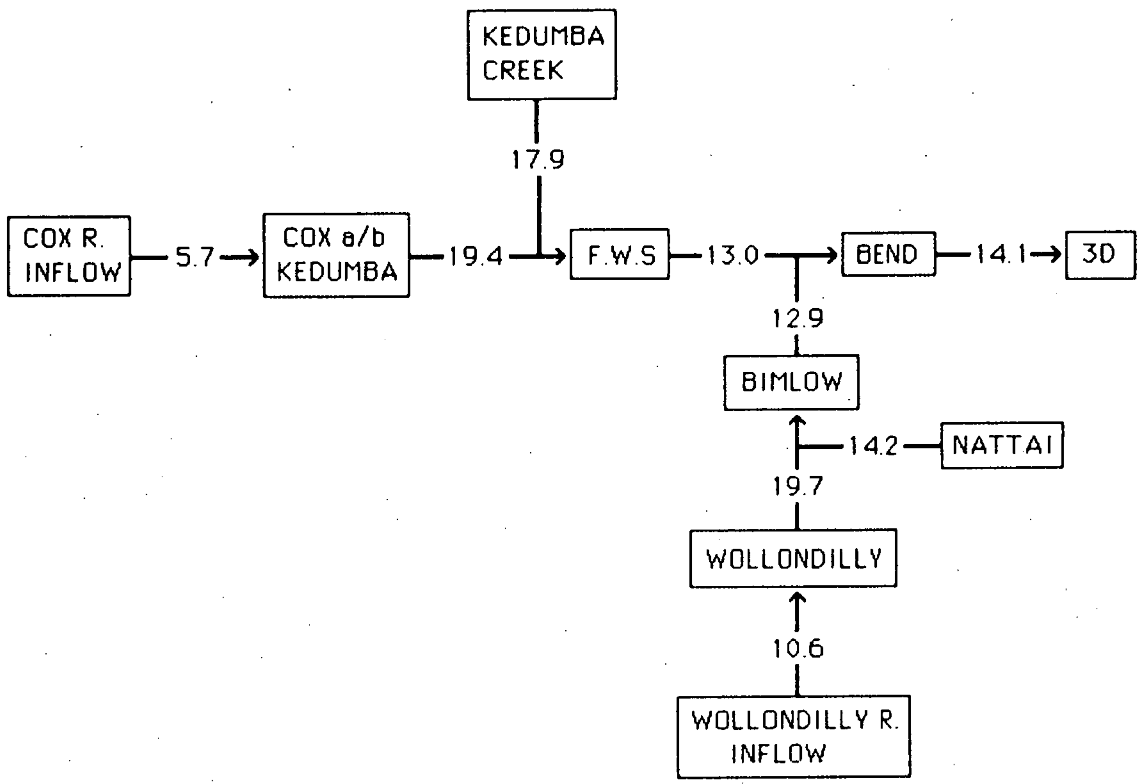

Cox R. above Kedumba Ck
Kedumba Ck
Warragamba Dam
Cox R.
inflow
Flood Warning Station
[F.W.S.]
3D
Bend
Bimlow
Nattai
N
10 km
Wollondilly
Wollondilly R. inflow

FIGURE 2.2

A schematic representation of the sample sites in Lake Burragorang, showing the maximum distances (km) between them, measured along the old river course. That is the distance that would be traversed in the lake, by underflows. Distances are those between boxes, so that the distance from Nattai to Bimlow is 14.2 km and the distance between Wollondilly and Bimlow is 19.7 km. The position at which the Nattai to Bimlow and Wollondilly to Bimlow lines join gives no indication of the distance between the junction of the old Nattai and Wollondilly rivers and Bimlow. It simply indicates that the junction occurs between these sites.


KEDUMBA CREEK
17.9
COX R. INFLOW
5.7
COX a/b KEDUMBA
19.4
F.W.S
13.0
BEND
14.1
3D
12.9
BIMLOW
14.2
NATTAI
19.7
WOLLONDILLY
10.6
WOLLONDILLY R. INFLOW

for safety reasons, the profiles taken from Warragamba Dam itself are restricted to the upper 60 m. This results in a loss of information on the water quality of underflows in particular.

Considerably more detailed profiles of temperature (1 m intervals to 30 m, and 3 m intervals thereafter) have been collected by the Biology Section personnel. These profiles are not considered here because none of the other profiles have a similarly detailed depth resolution, but they would facilitate a more exact study of vertical heat diffusion than is presented here.

Physical and Chemical Measurements

Inevitably with a long term study, methods have been changed to accomodate changing technology and scientific perception of the basic parameters required for the management of water supply. In most instances method changes are not considered to have altered the accuracy of the measurements. Exceptions to this are detailed below.

Temperature (°C):-

Profiles were measured using a thermistor and Wheatstone bridge circuitry. In earlier years a reversing thermometer was used for sampling at depths below about 60 m. Within a single profile the accuracy is ± 0.05°C. Between different sample days the accuracy is generally about ± 0.1°C, judged from temperatures taken at the deepest sampled depth, but in 1969 a regular fluctuation of about 0.7°C occurred in the samples from 82 m indicating that at worst ± 0.4°C might apply. This sort of variability is reduced by averaging to monthly figures.

Water sampling:-

Within the study period Van Dorn, Friedinger, E kman, Ruttner and Kemmerer discrete depth samplers, ranging in capacity from 1 - 5 litres, have been used. Integrated samples (0 - 4.5 m) taken with a Lund tube were designed to approximate the mean euphotic depth in the lake, though Secchi transparencies for site 3D indicate that the euphotic depth can be greater than 4.5 m, and in April and May 1980 euphotic depths of 12 m - 14 m were accurately determined using a Li-Cor submersible quanta meter.

Dissolved Oxygen (mg l^{-1}):-

Determined using the Winkler method, with azide modification, following standard methods (see Anon. 1971). Percent saturation was calculated using equations (after Thomas 1979) which reproduced the data of Green and Carritt (1967). No altitude correction was employed (the surface of Lake Burragorang lies about 117 m above sea level), and salinity was assumed equal to 0 gm kg^{-1}.

Turbidity (Hellige units):-

A Hellige visual comparator was used for the period prior to September 1979. The data reported here does not include any measurements by the new method, although the estimation of turbidity for late 1979 and 1980 (using the linear relationship between turbidity and total iron, mg l^{-1}; see Chapter 6) indicates that the turbidity was less than 5 (old scale) during the latter part of 1979 and in 1980, throughout the water column at site 3D.

Total Iron (mg l^{-1}):-

Prior to 1971, total iron was determined using a visual, colour disc comparator system, and results were reported to an accuracy of 0.05 mg l^{-1}. After early 1971, spectrophotometric colour determination was used, and measurements reported to an accuracy of 0.01 mg l^{-1}.

Chloride (mg l^{-1}):-

Chloride concentration was originally determined using the Argentometric method but this was later changed to the Mercuric Nitrate method; both follow standard methods (see Anon 1971).

Alkalinity (mg l^{-1} as $CaCO_3$):-

Determined according to standard methods (see Anon 1971), using a colorimetric endpoint to the titration.

Nitrogen:- (mg m^{-3})

Total nitrogen concentration (mg m^{-3}) in samples from the variable offtake (0 - 60 m) at Warragamba Dam, is reported as albuminoid nitrogen plus ammonia and nitrate (nitrite was negligible). This underestimates total nitrogen (TN) and will be denoted here by tN. I. Smalls (pers. comm. 1984) reports that albuminoid nitrogen is about 60% of total Kjeldahl nitrogen for Prospect Reservoir, which receives much of its water from Lake Burragorang. Determination of albuminoid nitrogen and ammonia followed standard methods (Anon. 1971). Nitrate was determined by Nesslerization after reduction with hot alkaline Devarda's alloy. A significant method change occurred (for Nitrate) in October 1974, when Devarda's Alloy was introduced to replace Aluminium in the reduction process. Values of nitrate concentration increased to 5 - 7 times the pre-October levels. No intercalibration of the methods was attempted (A. Robertson pers. comm. 1982).

Phosphorus (mg m^{-3}):-

From January 1970, total phosphorus concentration has been determined for subsamples of the integrated (0 - 4.5 m) water samples placed in iodine-impregnated Polythene bottles and refrigerated immediately. Within 2 days the unfiltered samples (200 ml) were evaporated to c. 50 ml, then digested for 60 min with potassium persulphate, ammonium molybdate and potassium antimonyl tartrate. Following reduction with ascorbic acid and extraction with isobutanol, absorbances (690 nm) were compared with prepared standards, using a Varian 635D spectrophotometer. The method follows that of MacKay (1975). In 1979, phosphorus analyses were automated (Technicon AA2 auto-analyser), and two minor changes were made to the chemistry of total phosphorus determination. Ammonium persulphate replaced potassium persulphate in the digestion procedure, and hydrazine sulphate was added to the ascorbic acid reducer.

Chlorophyll-a (mg m^{-3}):-

Since January 1970, Chlorophyll-a concentration (mg m^{-3} total pigment uncorrected for phaeopigments) has been determined, from subsamples of the integrated (0 - 4.5 m) water samples, by the methods of the Scor-Unesco Working Group No. 17 (1966). After January 1980, the trichromatic equations of Jeffrey and Humphrey (1975) were used to estimate chlorophyll-a.

Secchi transparency (m):-

Determined using a standard disc with black and white quadrants, and without the aid of a water telescope so that water surface roughness and reflection are not precluded.

DATA ANALYSIS AND REPRESENTATION

Storage and graphics

Data storage was on floppy disk using a Hewlett Packard 9825A desk-top computer. Two basic file structures, a depth versus time file for physical and chemical profiles, and a parameter versus time file for individual parameters, were developed for the data and interfaced with small programs for univariate statistics and simple bivariate plotting (e.g. parameter/depth or time plots). Plotting, on an HP 9862A platen plotter, employed graphics subroutines developed for the HP 9825A by P. Minchin.

Data was transferred directly from the HP 9825A to the University main-frame computer, a Burroughs B6800, for isopleth representation, using the SURFACE-2 contouring graphics package (Sampson 1978).

Limitations of SURFACE-2 Contouring:-

SURFACE-2 has a number of intrinsic limitations, and others introduced by the modification of the data set to allow plotting relative to a fixed datum rather than to the water surface.

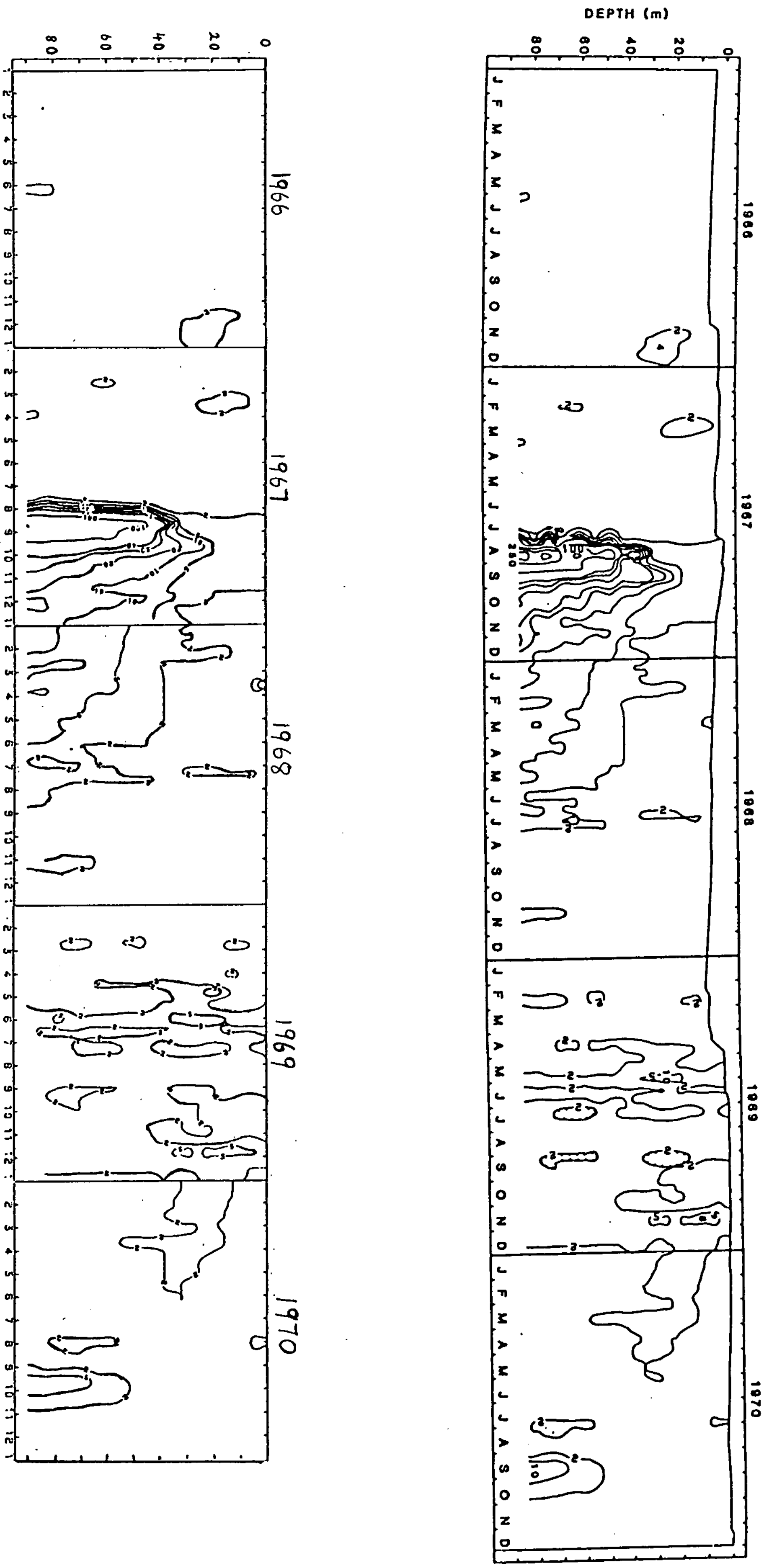
First, SURFACE-2 takes an irregular grid matrix of raw data and estimates a specified regular grid from a nearest neighbour search weighted to the squares of the distances from the point to be estimated. The practical effect is that instantaneous events, such as a massive turbid inflow, are smoothed out along the time axis, appearing as a gradual waxing rather than an abrupt event. The problem is less evident in the vertical plane. An example is the turbid inflow between 1/8/67 and 8/8/67 (Fig. 2.3, upper panel). From the plot it appears that the turbidity increase began in July. In fact, the first manifestation was on 8/8/67. The sparser the data the more acute the problem, and this is actually a feature of any attempt to represent a parameter using isopleths, because it arises from the use of linear interpolation between sample days, which is probably a greater approximation than linear interpolation between different depths in a single profile. A corollary of the above limitation is that SURFACE-2 will tend to reduce peak heights for the parameter being represented (ie temperature, turbidity etc.), a problem that is worst when sudden changes occur, as is the case for turbidity. To overcome this, the maximum turbidity recorded during the major inflows is marked on the turbidity isopleth diagram (see Fig. 2.3).

Second, there is a "bulging effect" where the vertical distribution of a parameter is plotted falsely, sinuously, as in the leading edge of the turbid inflow in the example cited above. Figure 2.3 (lower panel) shows a plot of the same data, without the modification allowing for surface level fluctuations. The bulging problem is not evident, but the horizontal "spreading", in time, remains.

It is emphasized that these limitations appear principally when sudden changes take place. Despite such limitations, which if understood are only minor irritants, the value of the SURFACE-2 plots is that they give a concise summary of the behaviour of the reservoir and major events in its history. They should not be used to provide accurate individual profiles for single sample dates.

FIGURE 2.3

Isopleths of turbidity (1966 - 1970) drawn by SURFACE-2 (R. J. Sampson 1975, revised 1978) showing the effect of adjusting each profile to a fixed datum on the dam wall (Full Supply Level, FSL). The upper panel is drawn relative to FSL, and shows the "bulging effect" at points where the turbidity increases rapidly (ie July/August 1967), while the lower panel is drawn relative to the lake surface and has no bulging problem.


DEPTH (m)
1966
1967
1968
1969
1970

Long term averaged profiles

Profiles of monthly mean temperature and dissolved oxygen (1961 - 1980) were calculated in two stages.

First, for each of the twenty years and at each of the 11 standard sample depths, a monthly mean was calculated, giving 12 profiles (1 $month^{-1}$) for any one year. This step removed the effect of changing sampling frequency by yielding a single profile for each month of each year in the study period.

Second, at each of the standard depths a grand mean (n = 20 usually) was calculated from the individual monthly means, yielding a single profile for each month over the full study period. Consequently, the standard deviation for any point (ie January, at 6 m) describes the dispersion of the 20 individual monthly means (one for each year 1961 - 1980) about the overall mean.

The 11 standard depths referred to above are not precise as they usually represent a small range of sample depths. This is not considered significant except for the deepest sample (nominally 82 m), which has ranged from 48.8 m to 91.4 m. However, examination of the frequency distribution of the deepest sample, for the full study period, shows that 48% of all samples were from 82 m and 86% from below 75 m.

It should be noted that the "metalimnion" of these 20 year average monthly profiles is a zone of great variability, and the gradient of the averaged profile may not reflect the metalimnion found on any of the sample days from which it was derived. The average profiles could result from a series of profiles with similar, comparatively gentle gradients, or from a series of profiles with sharp, near horizontal gradients, occurring at a variety of depths. A combination of both extremes is most probable. It can be said with certainty that the epilimnion of the long-term average profile represents the minimum vertical extent of the epilimnia of its constituent profiles, and similarly for the near isothermal hypolimnion; but the metalimnion is a region of doubt which may be either epilimnetic or hypolimnetic in character or it may accurately reflect the gradients of its constituent metalimnia.

For this reason, a series of representative profiles of temperature and dissolved oxygen (one month^{-1} for 8 of the 20 years) is given in Appendix 2.

Interestingly, it may be argued that the basic nature of the metalimnion is that of a region containing the accumulated images of thermal profiles having near horizontal separation between the mixed and unmixed layers. Consequently, the metalimnetic gradient is relatively gentle compared to the boundary between mixed and unmixed layers at any instant. Perhaps a more realistic view of the metalimnion, would be that of a region in which the probability of mixing is declining relative to that within the epilimnion. Again, this can also be said of the long term averaged profile.

Calculation of Heat Content, Stability, and Birgean Wind Work

A version of LIMNO, written by D. H. Merritt (see Johnson et al 1978) and modified by J. Ferris (see Appendix 1), was implemented on the Burroughs B6800 computer and used to estimate whole-lake mechanical stability, heat budgets and chemical budgets.

Calculation of Inflow Volumes

Early in this study the cardinal role of inflows in determining overall water quality and affecting some aspects of stratification behaviour became apparent. It was therefore necessary to have a complete record of total monthly inflows. An estimate of total monthly inflow volume was made by calculation from the following data:-

1. The total discharge to Prospect Reservoir, and through the hydro-electric offtake per day.
2. The instantaneous storage level at 8.00 a.m. each day (Relative level; RL).

3. The instantaneous flow rate, in times of flood, over the dam crest or through the radial gates. The flow rate is assumed to hold from 8.00 a.m. Day 1 to 8.00 a.m. Day 2, etc.

The total monthly inflow (minus evaporation) is calculated as total monthly discharge + flood loss ± the gross change in storage volume between first and last days of each month (calculated using daily records of RL and hypsometric data).

The weakest aspect of this calculation is the assumption of constant RL for successive 24 hour periods. Data from the detailed records of flood RL's (hourly) has not been incorporated in the inflow estimates. Nevertheless the data presented here can be taken to provide one of the best estimates of total inflow into Lake Burragorang, and the record is very nearly complete over the full 20 years studied. The records of inflow from direct stream gauging are not complete for the study period.

A Register of Effective Inflows

The "Inflow Register" (Table 3.3) is a listing of all inflows which have had significant influence on physical and chemical conditions at site 3D. Recognition of the arrival of an inflow at site 3D was based primarily on changes in turbidity but correlated with other measured parameters such as chloride, iron, temperature, dissolved oxygen. This was, necessarily, a somewhat subjective procedure. The following criteria were applied:-

1. Any turbidity change should be > 2 units relative to all other measurements.
2. The change should appear at more than just the deepest sample point (i.e. avoid chance of contamination by mud disturbed by sampler).
3. A change in turbidity has more validity if it persists beyond a single sample day.
4. A change in turbidity should coincide in time and space with other

chemical effects, such as changes in iron or chloride.

5. The observed changes must bear some relationship to inflow/outflow records.

On these grounds, clear examples of inflows reaching 3D were recognized. There were borderline cases and cases of apparent sampling error or contamination. Only 37 cases were accepted for the Register out of 65 possibilities.

Linear Regression

All linear regression analysis employed TEDDYBEAR, developed by J.B. Wilson, Botany Department, University of Otago, and implemented on the University Burroughs B6800 computer. Confidence and prediction limits (95%) were calculated according to the method given by Sokal and Rohlf (1981).

WET AND DRY YEARS (CHAPTER 5)

Analysis

Two groups of 4 years were chosen to represent wet (1963, 1974, 1976 and 1978) and dry (1965, 1968, 1979 and 1980) conditions respectively. The monthly mean profiles and the differences between profiles from month to month were compared using a t-test which made the least restrictive assumptions concerning distribution about the mean (see Steel and Torrie, 1981; p 106). Using these averaged profiles, the heat content, Schmidt stability and Birgean wind-work were calculated for the two groups also. The use of a single monthly profile for each group (means for each depth; n = 4) in the calculation of these terms, rather than making a separate calculation for each of the four years, means there is no direct indication of the statistical significance of the differences between wet and dry groups.

Comments on the Analysis:-

There are significant difference between the two groups of years, in relation to the water level (and therefore the volume of water) in the lake. This biases whole lake parameters such as heat content and stability, without recourse to any immediate inflow effects and is therefore misleading in the context of the present analysis. In order to overcome this bias, the heat content, stability and Birgean wind-work (ie all parameters calculated using Limno) were re-calculated as though the lake were full. Although this ignores possible differences in heating processes between the two groups which may be dependent on surface area or volume, I consider it a justifiable manipulation in the attempt to discount the known volume dependence of the whole lake heat content and probably stability, and focus on the differences arising from inflow and outflow. This procedure is also consistent with the averaging of profiles which were measured relative to the water surface rather than a fixed datum. The importance, or otherwise, of this volume difference can be estimated from a comparison of the relevant quantities calculated first using mean monthly water levels and then assuming that the lake is full (Figs 2.4 and 2.5). The wet years undergo a relatively minor change in the first three months of the year (Fig. 2.4). The dry years, on the other hand, show increases in both heat content and Birgean wind-work that appear to be simply and directly related to the increase in volume, while the Schmidt stability is most affected during periods of maximum stability (Fig. 2.5).

The heat content of the dry years increases by up to 7% (for a volume change of up to 22% of the total lake volume, at FSL), while the heat budget is virtually unchanged (a fall of 1.4%) because the difference between maximum and minimum heat content is preserved almost unchanged. Schmidt stability increases with the increased volume by up to 25% in winter, this is not such a surprising result given the very small stabilities of the winter months when small differences constitute a greater percentage change. However, 14% is

FIGURE 2.4

Seasonal progression of heat content (cal cm^{-2}; x 4.19 = Joules cm^{-2}), Schmidt stability (gm-cm cm^{-2}; x 9.807 10^{-15} = Joules cm^{-2}), and Birgean wind work (gm-cm cm^{-2}) for WET years. The upper most line (□) is calculated assuming the lake to be full (at FSL), while the lower line (·) is calculated using the mean depth for each month. It is evident that in WET years the lake is full for most of the year and the assumption of a full lake makes little difference to the calculation.

Symbols:-

Calculated assuming FSL (□)

Calculated assuming mean depth (·)

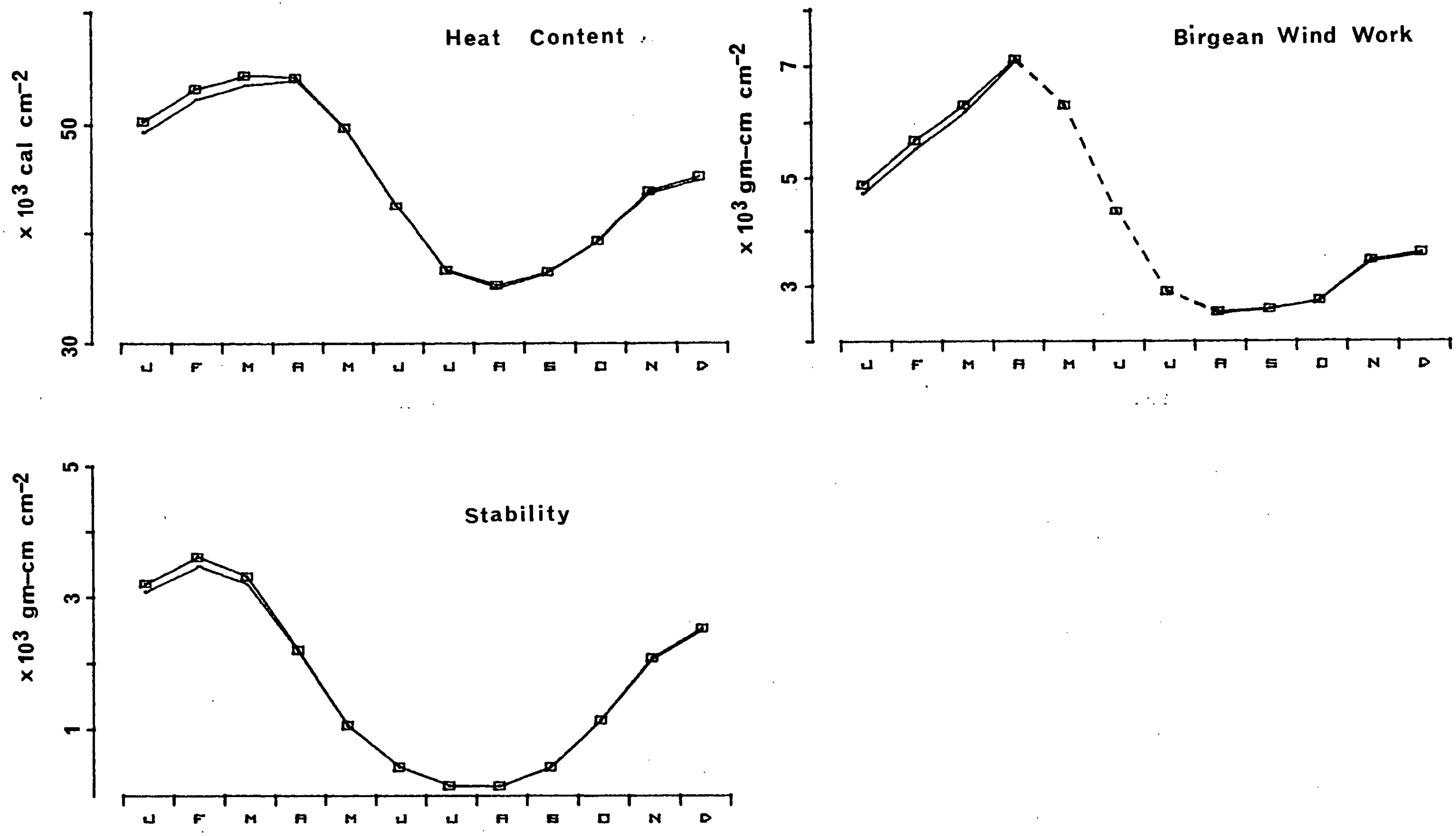

Heat Content
x 10^3 cal cm^-2
50
30
J F M A M J J A S O N D
Birgean Wind Work
x 10^3 gm-cm cm^-2
7
5
3
J F M A M J J A S O N D
Stability
x 10^3 gm-cm cm^-2
5
3
1
J F M A M J J A S O N D

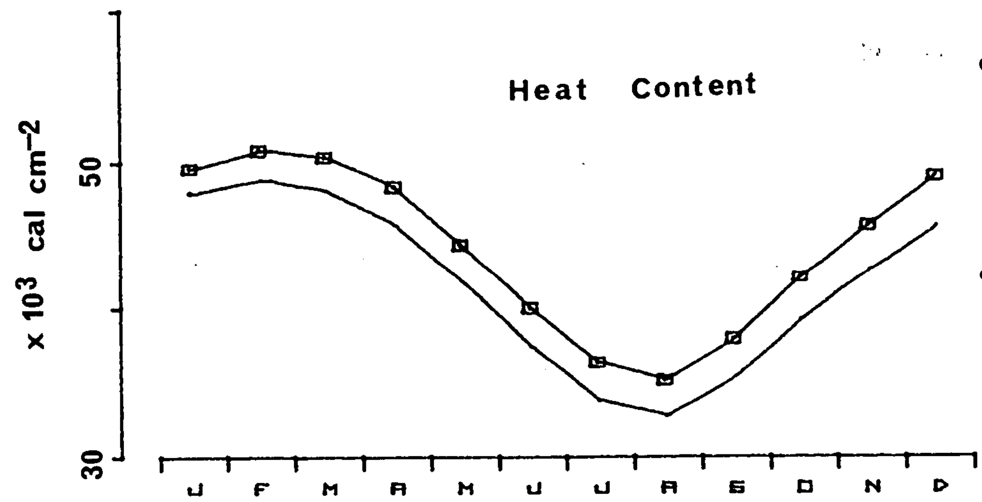
FIGURE 2.5

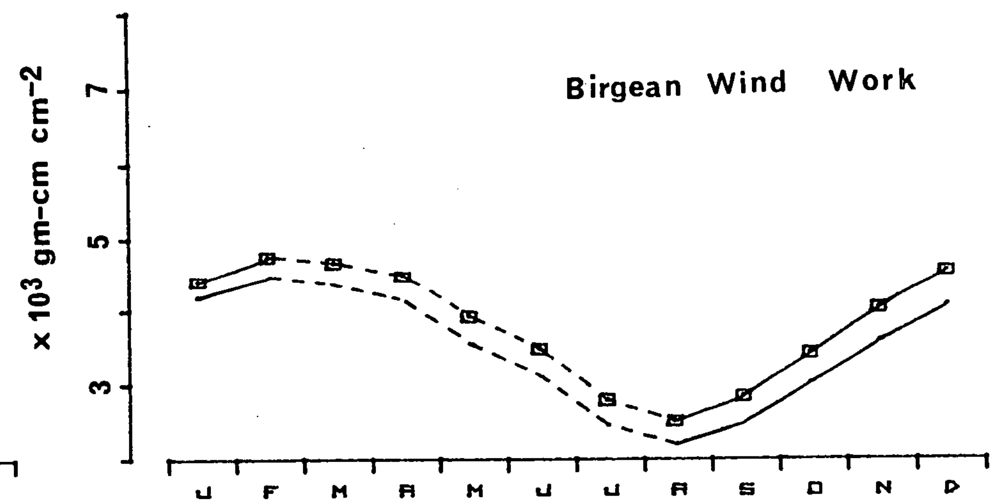
Seasonal progression of heat content (cal cm^{-2}), Schmidt stability (gm-cm cm^{-2}), and Birgean wind work (gm-cm cm^{-2}) for DRY years. The upper most line (□) is calculated assuming the lake to be full (at FSL), while the lower line (·) is calculated using the mean depth for each month. In DRY years the assumption of a full lake makes an appreciable difference to all three parameters. The difference is directly related to volume in the case of heat content and Birgean wind work, but a more complicated relationship exists for the stability.

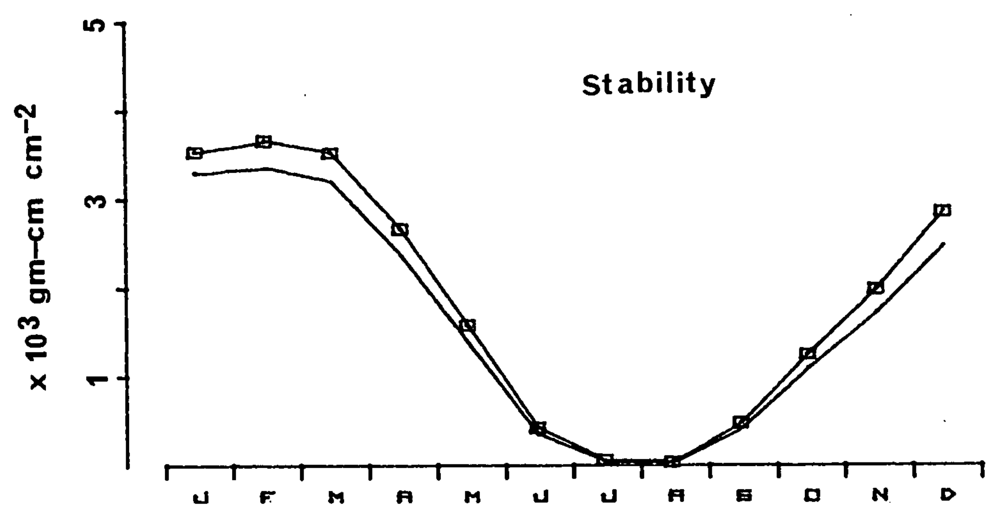
Symbols:-

Calculated assuming FSL (□)

Calculated assuming mean depth (·)


Heat Content
x 10^3 cal cm^{-2}
50
30
J F M A M J J A S O N D


Birgean Wind Work
x 10^3 gm-cm cm^{-2}
7
5
3
J F M A M J J A S O N D


Stability
x 10^3 gm-cm cm^{-2}
5
3
1
J F M A M J J A S O N D

probably a more representative figure for the data, over the full year. Changes of up to 12% are found for the Birgean wind work. Although these changes are a real aspect of the difference between wet and dry years in Lake Burragorang, they would clearly alter the interpretation of the present comparison, and are not directly related to the effect of inflows on the thermal structure in the water column. Further, it is apparent from Figs 2.4 and 2.5 that use of the mean monthly water levels would exaggerate the differences between wet and dry years. Consequently, the data reported in Chapter 5 represents the minimum differences between the two groups.

Both the annual heat budget and the Birgean wind work are sequential quantities, which in the Southern Hemisphere are usually calculated relative to the winter of the previous year. In the present treatment there is no data from a previous year, so that I have assumed circularity of the annual cycle. This is not strictly justified, given that the data groups consist of a small number ($n = 4$) of non-contiguous years, but the winter minimum heat content and Birgean wind work values are relatively conservative properties of the data set (varying by < 2% between WET and DRY years, and the 20 year mean data), so that the assumption of circularity does not entail significant error.

Following from this, it is apparent that January differs from any of the other months insofar as the previous month (December) may have been wet or dry. Consequently, the January comparison may have one less component of the variation between WET and DRY years and show only those effects occurring within the month. In fact, the December preceeding the WET years has a mean inflow (minus evaporation) of 53 10^6 m^3, compared to 4 10^6 m^3 for the December preceeding the DRY years. January of the WET year group does, therefore, have a greater inflow in the preceeding month, but the volume is less than any of the WET year monthly means.

Comments on the Wet and Dry Years

A number of features of the chosen years are relevant to the full understanding of this comparison and are listed (with comments) below:-

1. The years were chosen, as far as possible, to be consistently wet or dry, rather than having aspects of both. For example in 1964 in which a single very large inflow (in June) accounted for much of the annual inflow total, and 1966 a single flood event (in November) occurred in an otherwise dry year. In this context 1976 is the least satisfactory of the wet years, because about five months were comparatively dry.

2. The wet years (1963, 1974, 1976, 1978) had total annual inflows (minus evaporation) ranging from 2074 - 3904 10^6 m^3, yielding retention times less than 1 year (0.53 - 0.99 years). The dry years (1965, 1968, 1979, 1980) had inflows totalling 259 10^6 m^3, or less, giving retention times from 7.9 - 22.8 years. The comparison is, therefore, between data blocks that differ in annual inflow (minus evaporation) by about an order of magnitude.

3. Although there is greater inflow for every month of the WET years than for corresponding DRY months, there a few WET months with relatively low inflow averages, particularly February, November and December. Consequently, these months may be expected not to reflect the direct influence of inflows to the same extent as some of the other months, such as March - June.

4. The operation of the outflow structures, differs markedly between the two groups, in that the HEPS offtake is nearly unused for the DRY years (except for the first few months of 1965), while it is in almost continuous use during the WET years (except for May 1976, and December of 1974, 1976, and 1978). The use of water supply offtake,

also differs slightly, being generally lower in WET years presumably because the water quality is lower in WET years and the offtake is closed more of the time. However, this difference in volume between the two groups is insignificant relative to the large differences in inflow volume and the volume subtracted through the HEPS, and can be ignored.

5. Finally, it is interesting to consider that the chosen DRY years (1965, 1968, 1979, 1980) are all characterised by a convective overturn in winter, a total lack of surface outflow, and none is represented in the register of effective inflows. In comparison, the WET years (1963, 1974, 1976, 1978) all show advective stabilisation in winter, commonly exhibit surface outflow (flooding over the dam-crest), and have at least one registered inflow.

TOTAL PHOSPHORUS - CHLOROPHYLL (CHAPTER 7)

Regression analyses used TEDDYBEAR, an ANOVA correlation and regression package by J. B. Wilson, Otago University. It routinely tests residuals for skewness, kurtosis and correlation between absolute magnitudes of residuals and the independent variable. For conformity with the linear regression model it was usually necessary to transform the data to logarithms, so for uniformity, ln transformation was applied throughout. Because re-transformation of double-ln data to arithmetic values estimates median values rather than means (of the original distribution), when retrieving arithmetic values we applied a multiplicative correction factor (F) (Baskerville 1972; Sprugel 1983). This correction factor is calculated from the mean square of deviations ($s^2_{y.x}$: residual mean square) from the regression line as follows:-

$$F = \exp(s^2_{y.x}/2)$$

The corrected predicted value of chlorphyll-a is thus given by the equation

$$C = (\exp y) \times F$$

where C is chlorophyll-a concentration (mg m^{-3}), y is the predicted concentration of chlorophyll-a obtained from the regression equation. Correction factors are reported for all regressions, including the re-analyses of published data. Statistical methods for determination of 95% confidence intervals and 95% prediction limits (the latter apply to prediction of the dependent variable from a single value of the independent variable) follow Sokal and Rohlf (1981). The correction factor is not applicable to the recovery of arithmetic values of the prediction limits (V. Smith pers. comm. 1984).

The regressions reported here use two primary variables, concentrations of total phophorus and chlorophyll-a. The various statistics that may be derived from regular measurements of these variables (annual mean, seasonal mean, annual maximum, seasonal maximum, etc.) are collectively referred to here as subvariables.

PART II

Chapters 3 - 5

INTRODUCTION

In this part of the thesis, stratification cycles of both temperature and oxygen are examined. Initially, the account of Lake Burragorang's behaviour is based on a condensation of the full twenty year data set in the form of long-term mean monthly profiles. These data, which inevitably ignore much of the shorter term variation and the processes that generate that variation, is used to place Lake Burragorang in the context of others lakes in Australia and elsewhere. In the next stage, the range of year to year variation is examined more closely and the factors that potentially contribute to that variation are identified. Perhaps not surprisingly, inflow and outflow (collectively, advective effects) appear to exert a profound influence on the important physical cycles in the lake, and to generate a range of behaviour commensurate with their own variability; in a region without pronounced seasonality of rainfall and historically subject to prolonged periods of drought. Arising from this, an analytical experiment designed to test the different thermal and oxygen stratification of wet as opposed to dry years is described. This experiment is based on the premise that Lake Burragorang's behaviour, in drought periods, approaches that of a natural lake and can be substantially accounted for in terms of classical limnological theory, while the cycle found in wet years demonstrates the effects of advection and can contribute to the development of limnological theory in a field that is poorly researched, especially in this country.

CHAPTER 3

THERMAL STRATIFICATION

DESCRIPTIVE ACCOUNT OF THERMAL STRATIFICATION

Warm Monomixis

Consideration of Lake Burragorang's geographic position (Latitude 33° 55' S: Altitude 116.7 m above sea level) places it in the region typified by warm monomictic lakes, according to the classification of Hutchinson and Loffler (1956; after Wetzel 1983). Such lakes are thermally stratified in summer, followed by a single period of winter mixing, when water temperatures remain above 4°C (Hutchinson 1957). Lake Burragorang is situated near to the N.S.W. coast (approximately 50 km West of Sydney). The lake is, therefore, influenced by an oceanic climate which further pre-disposes it to a warm monomictic thermal regime. Hutchinson (1957) observes that another effect of the oceanic climate is to blur the usually sharp tri-partite thermal stratification typical of lakes influenced by a more continental climate.

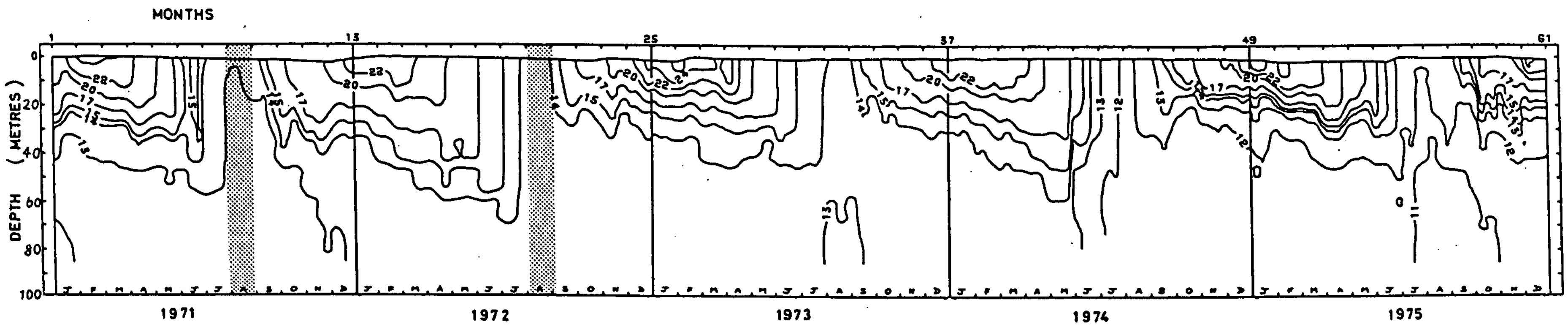
The thermal stratification of Lake Burragorang is illustrated, for the study period (1961 - 1980), in Fig. 3.1. The data is that of the Warragamba Chemistry Laboratory (MWS&DB) and is plotted relative to a fixed datum (Full Supply Level; FSL) so as to be horizontally consistent with the offtake structures. Fluctuations of the lake surface (Fig. 3.1) are indicative of the lake's water balance. A depth of 0 m equals FSL; any further rise results in flooding directly over the dam crest. Two extended periods of drought are evident in 1968 - 1969, and again in 1979 - 1980. In the latter period Lake Burragorang was drawn down to 11 m below FSL, the lowest record since November 1961 when the lake first filled.

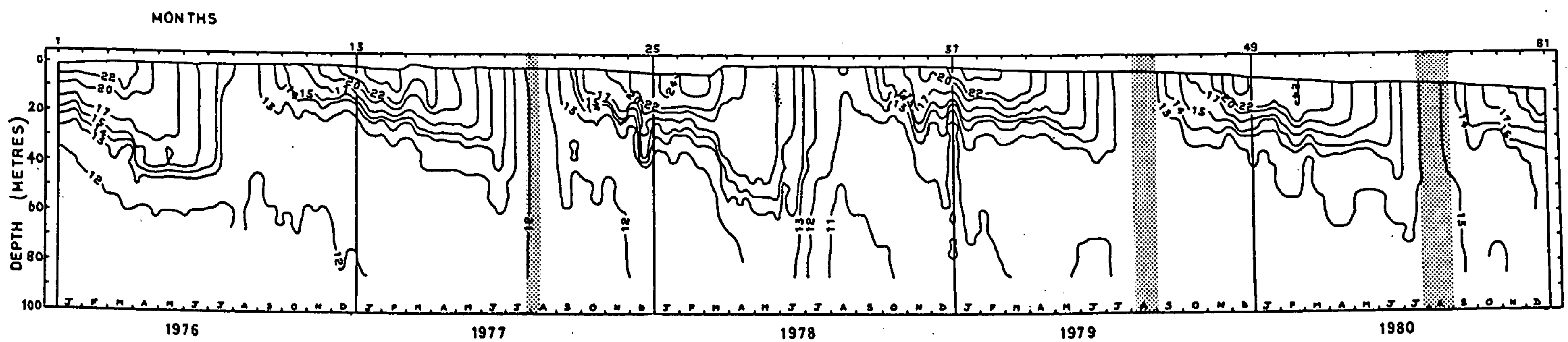
From Fig. 3.1 it is apparent that Lake Burragorang's cycle of thermal stratification conforms, basically, with the anticipated pattern for a warm monomictic lake. Thermal stratification is seen to last for 10 or more months of each year, leaving a relatively short period of near isothermy and potential

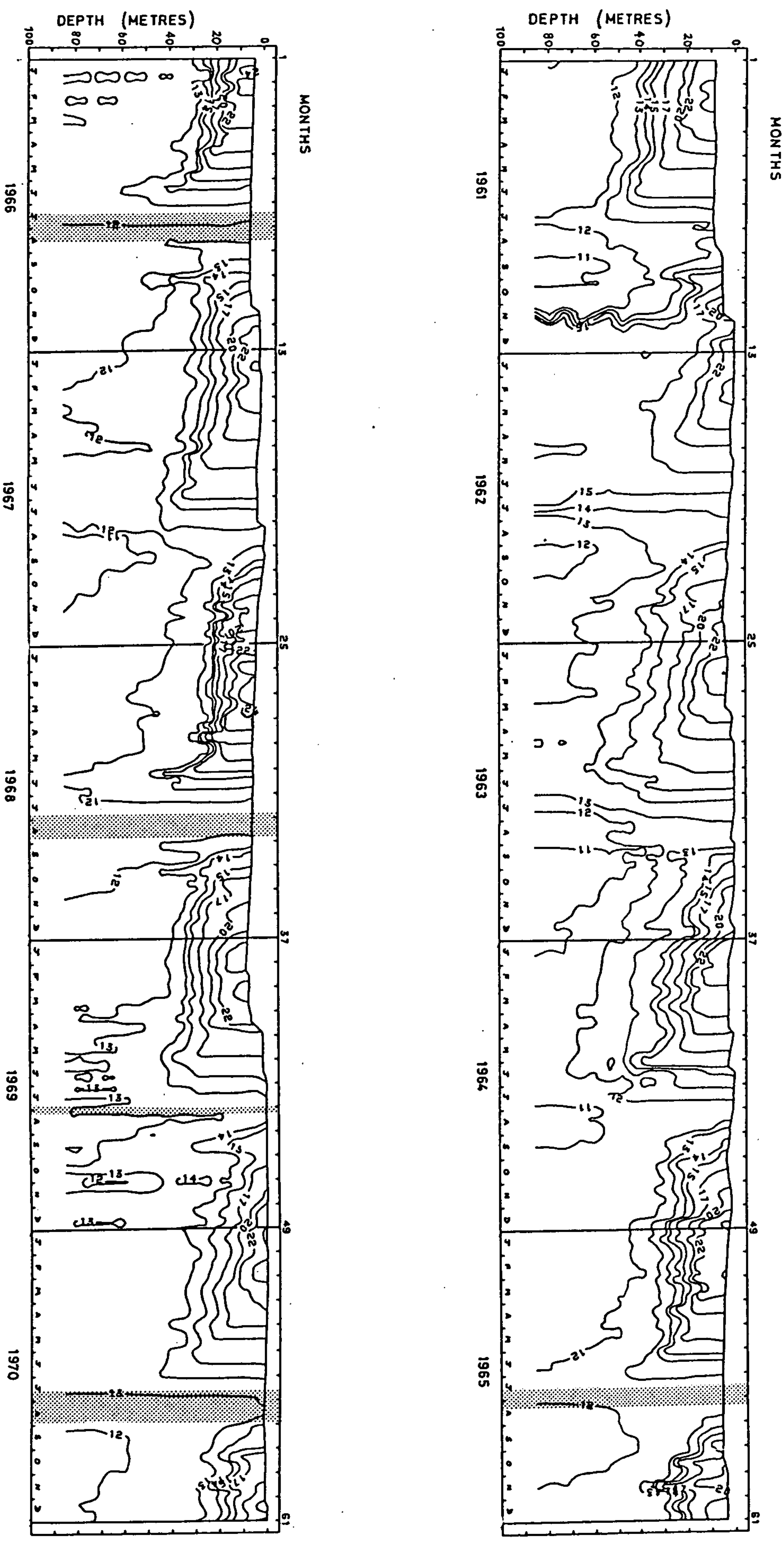
<u>FIGURE 3.1</u>

Temperature (°C) isopleth diagram, 1961 - 1980. Isopleths are drawn at the following temperature intervals:- 11, 12, 13, 14, 15, 17, 20, 22, 24, 26°C. Vertical lines are drawn at the end of each calendar year. Stippled areas represent approximate periods of overturn (ie circulation to the deepest sampled depth) subjectively estimated from profiles of temperature, dissolved oxygen, turbidity, chloride, pH, alkalinity, total iron, and requiring a near complete chemical and physical homogeneity of the water column, with no obvious discontinuity at any depth.

I have not used any arithmetic definition for the existence of stratification, like Birge's classic 1.0°C m^{-1}, or any of the similarly convenient but arbitrary definitions, for example Coche (1968) used 0.2°C m^{-1} for tropical Lake Kariba, partly because any such definition is dependent on the depth frequency of sampling (a 6 m, or 12 m depth interval between samples does not give a meaningful resolution for such a calculation), and partly because, as pointed out by Lewis (1973), the stability of a thermocline increases with its distance from the surface. In July 1966, a deep temperature gradient of 0.4°C (between 48 m and 61 m below the surface) was associated with an oxycline of 5.3 mg l^{-1} (50% of saturation) at site 3D in Lake Burragorang.


MONTHS
DEPTH (METRES)
1971
1972
1973
1974
1975


MONTHS
DEPTH (METRES)
1976
1977
1978
1979
1980



DEPTH (METRES)
MONTHS
1961
1962
1963
1964
1965
DEPTH (METRES)
MONTHS
1966
1967
1968
1969
1970

holomixis (1 - 2 months). A stratification period of about 10 months is comparable to other deep (> c. 50 m) reservoirs in Australia, such as Lake Barrington (Tasmania; Tyler and Buckney 1974), Lake Argyle (Western Australia; Imberger and Patterson 1979), and Lake Eildon (Victoria; Powling 1980). In the last instance, Powling (1980) lists a 7 month stratification period for Lake Eildon, but the data presented indicates that 10 months is more likely. Shallower reservoirs exhibit a shorter period of stratification. For example, Lake Hume (Victoria; 41.4 m) stratifies for about 5 months from December to April (Croome 1980), and North Pine Dam (Queensland; 35.1 m) is stratified for about 7 months from October to April (King and Everson 1980). Deeper reservoirs such as Lake Gordon (Tasmania; 140 m; Steane and Tyler 1982) and Dartmouth Reservoir (Victoria; 171 m; Welsh 1984) have remained permenantly stratified, although this is based on investigation of only two years in the case of Lake Gordon.

Holomixis does not occur in Lake Burragorang each year and this will be more fully discussed later. This does not, however, change the fact that Lake Burragorang is essentially a warm monomictic lake. Holomictic periods, when the lake has circulated to the deepest sampled depth at site 3D (adjacent to Warragamba Dam), have been determined from thermal and chemical profiles and are marked on Fig. 3.1. These periods occur in July and August, with water temperatures ranging from about 11 - 14°C. Throughout holomixis, cooling of the water column is never more than about 1°C. This, and the small range of temperature at the deepest sampled depth suggest that Lake Burragorang can be regarded as a first class lake, one for which increased depth would have little effect on its ability to store heat (cf Hutchinson 1957). At the peak of stratification (January - February) maximum recorded surface temperatures lie between 24.1 and 28.3°C, with most falling in the range 24 - 26°C. A vertical gradient of 12 - 15°C is common.

Seasonal Progression

Fig. 3.2 shows the seasonal progression of temperature at each depth. In Fig. 3.3 the mean profiles (with standard deviation) are plotted for each month, derived from the full 20 years of data. Fig. 3.4 shows the mean temperature changes at each depth, between months, for the study period. Bearing in mind the effects of such long-term averaging (see Methods, Chapter 2), it is possible to give the general outline of the thermal cycle in Lake Burragorang from Figs 3.2, 3.3, and 3.4.

Heating begins in September and in the period from September to December is greatest at the surface. The rate of change of temperature is greatest between September and October, when surface temperature increases by > 3°C (Figs. 3.2 and 3.4). In this period (September - December) there is no clearly defined epilimnion. This may simply indicate that the epilimnion is less than 6 m deep (ie not detectable from this data), or it may be a feature of the oceanic thermal profile. Steane and Tyler (1982) found that Lake Gordon lacked a definite epilimnion during the heating phase. Hutchinson (1957) remarks that British workers were misled by this phenomenon into believing the tri-partite structure of a stratified lake was a feature of autumnal cooling. Evidently the September to December period is characterised by heat input at the surface in excess of the rate at which it is distributed downwards to the next sampled depth (6 m). Nevertheless, Fig. 3.4 shows that the upper 20 m (approximately) is the zone in which turbulent heat transport is operating on average.

The beginning of a second phase is apparent from the December - January temperature difference plot (Fig. 3.4). For the first time the surface temperature increase is exceeded by the increase in the 6 - 12 m region. The rate of surface heating has fallen below the rate at which wind-induced turbulence is acting to distribute heat downwards. As the January temperature profile indicates (Fig. 3.3), this leads to the establishment of a

FIGURE 3.2

Seasonal progression of twenty year monthly mean temperatures (°C) for 11 standard depths (see Materials and Methods, Chapter 2). The same data is presented as profiles with standard deviations in the Fig. 3.3. This figure also gives more computational detail.

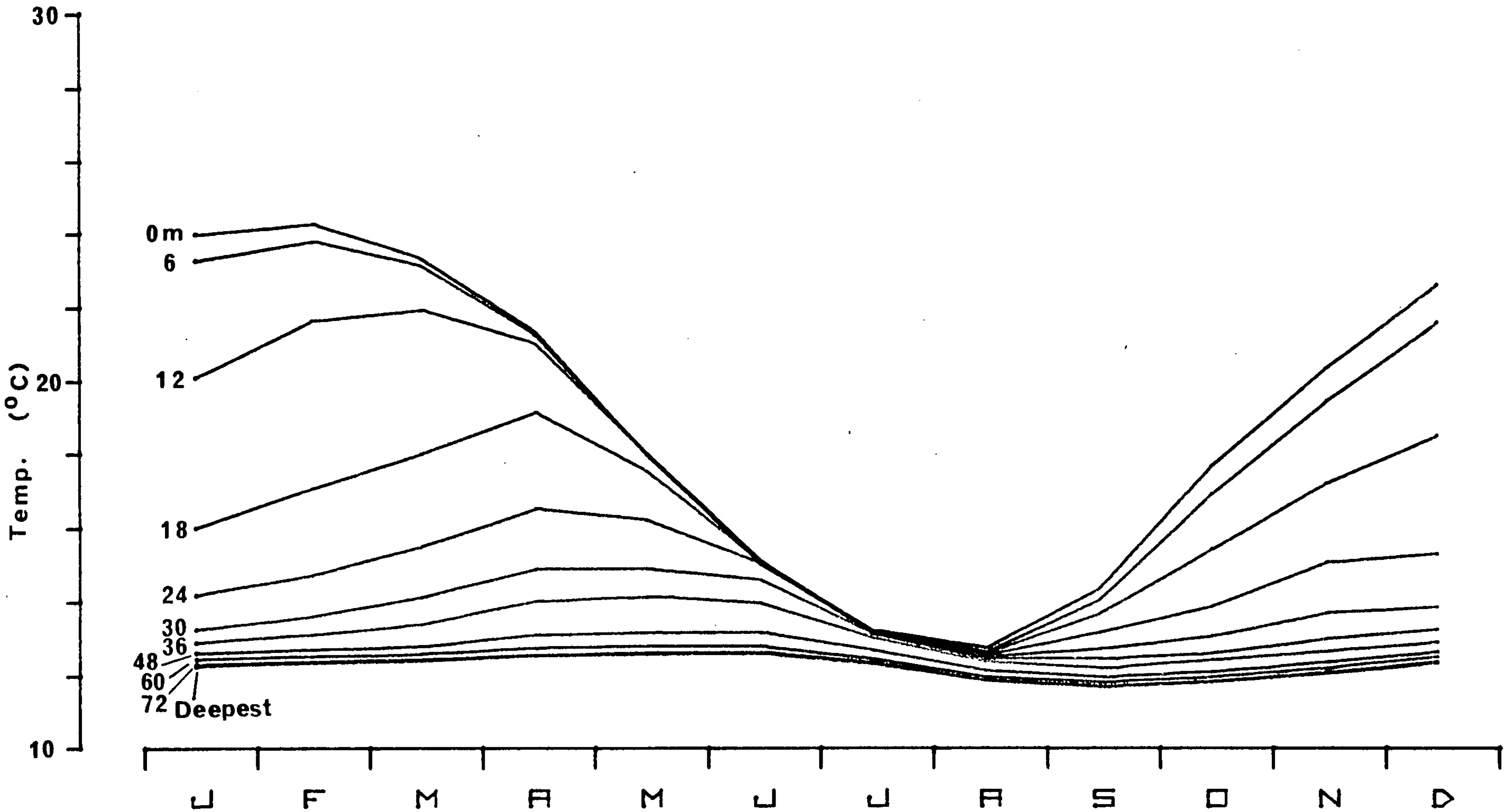
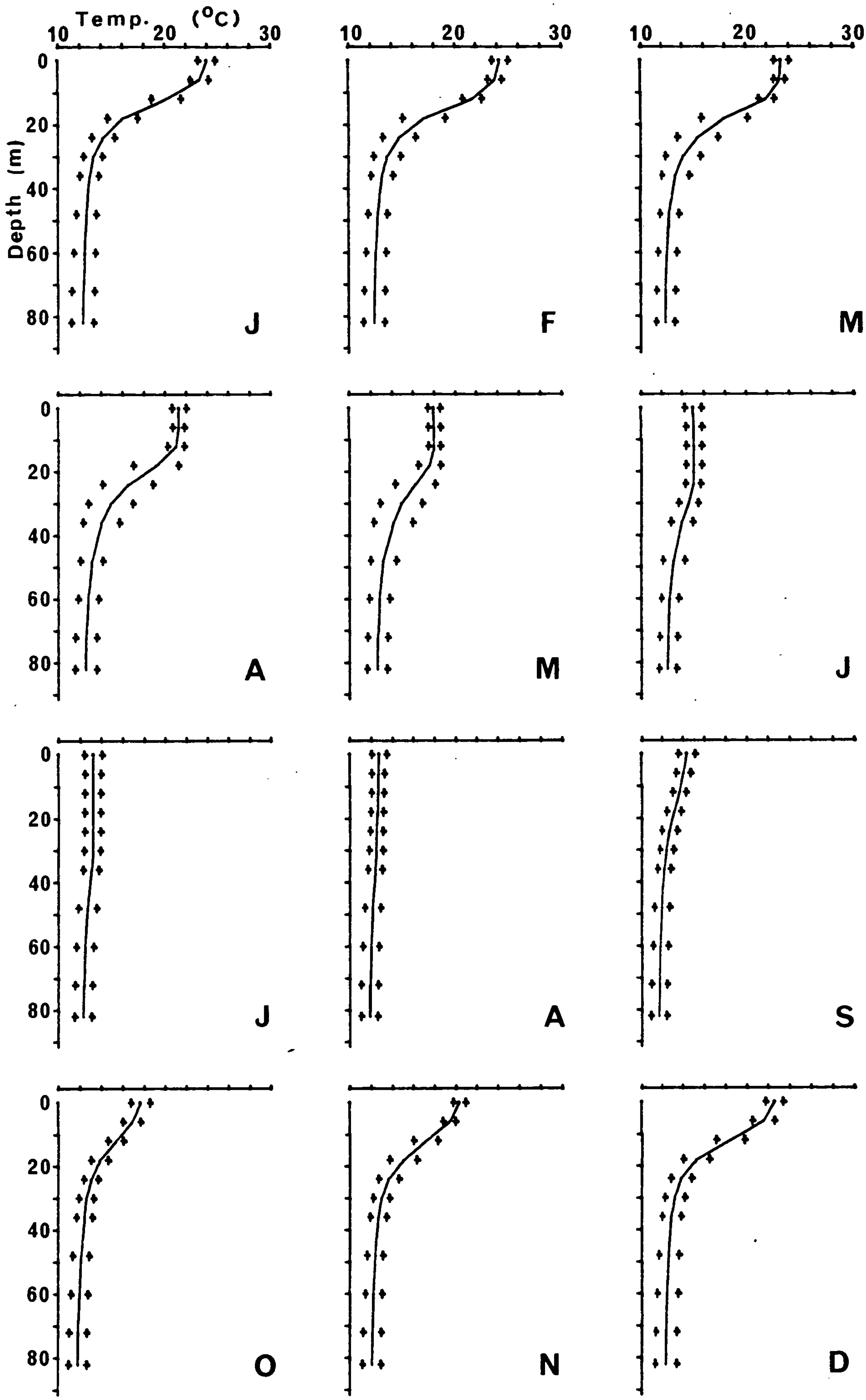

30
20
10
Temp. (°C)
0m
6
12
18
24
30
36
48
60
72
Deepest
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 3.3

Profiles of monthly mean temperature (1961 - 1980), calculated in two stages:- (see Materials and Methods, Chapter 2)

1. The monthly mean temperature, for a particular depth below the water surface (eleven depths in all), is calculated for each of the twenty years, yielding 12 (months) x 20 (years) x 11 (depths) means, each with an n of from 1 - 5.
2. Then, the mean of the 20 individual monthly means is calculated, yielding 12 profiles, each with 11 depths.

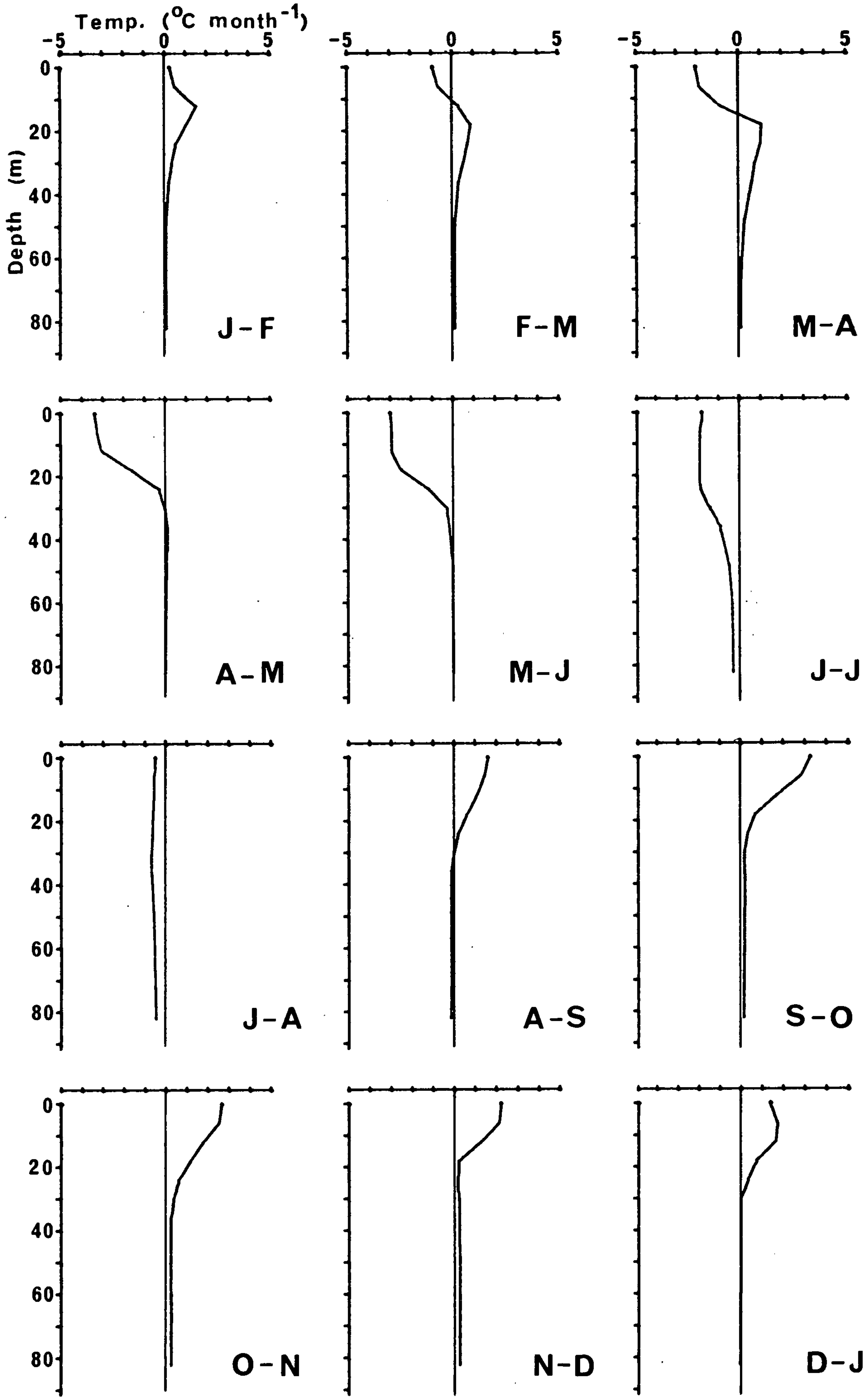
The first of these steps, is designed to remove the effect of changing sampling frequency, so that the weekly sampling of the early years do not overshadow the 2 weekly or monthly samples of later years. The standard deviation for any point (ie January, at 6 m) describes the dispersion of the 20 individual monthly means (one for each year 1961 - 1980), about the overall mean.



Temp. (°C)
10
20
30
Depth (m)
0
20
40
60
80
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 3.4

Profiles of monthly mean temperature difference (1961 - 1980) corresponding to the temperature profiles given in Fig. 3.3. These differences are calculated in an analagous manner to the monthly mean temperatures. The differences between months are accumulated for each of the 20 years, and then a grand mean of these differences is calculated. These plots demonstrate the dynamic aspects of temperature change at each depth. Johnson et al (1978), and Johnson and Merritt (1979) use the Birgean direct work curves in a similar manner. The plots presented here do not take into account the volume distribution with depth in the lake.


Temp. (°C month⁻¹)
-5
0
5
Depth (m)
0
20
40
60
80
J-F
F-M
M-A
A-M
M-J
J-J
J-A
A-S
S-O
O-N
N-D
D-J

recognisable epilimnion. There is still a net gain of heat at the surface until February.

Initiation of a third phase is evident from the February - March temperature difference plot (Fig. 3.4). Here, the surface temperature change is negative marking the onset of autumnal cooling. The epilimnion is progressively deepened with a concurrent deepening of the zone from which heat is being lost (Fig. 3.4). At first, this loss is (February - March, Fig. 3.4 simply a downward redistribution of heat rather than active cooling of the water column (there is no loss of total lake heat content); by March - April (Fig. 3.4) heat is being lost to the atmosphere as well as to the deeper layers. This re-distribution of heat appears as a gradually downward propagating wave of positive temperature difference between about 10 m (December - January) and 40 m (April - May, Fig. 3.2 and Fig. 3.4). Johnson et al (1978) compared Birgean direct work curves for five North American lakes, taking the month to month differences in a similar manner to the temperature difference plots presented here (Fig. 3.4) but taking into account volume distribution with depth. They found an analagous wave-like deepening of the region of maximum change in work for Mirror, Lawrence and Findley Lakes (see also Johnson et al 1984).

The May - June temperature difference plot (Fig. 3.4) indicates that temperature change is negative or close to zero throughout the water column and precedes the fourth and final phase, between June and August, when active heat loss affects the whole water column (Fig. 3.4). Between June and July heat loss is greatest in the upper 30 - 40 m of the lake, indicating that thermal stratification still exists (Fig. 3.4). This is also apparent from the mean temperature profiles (Fig. 3.3), though the sharpness of the thermocline is obscured by averaging. The cooling period in Lake Burragorang is usually characterised by much sharper gradients but at different depths from year to year. Consequently the overall mean profile has a minimized gradient.

Between July and August temperature change is nearly uniform with depth and approximately 0.5°C month^{-1}. The August mean temperature profile is nearly isothermal at about 12°C.

In summary then, the basic division of the annual cycle into a spring - summer heating period (September - February) and a cooling phase of 6 months in autumn and winter (March - August) is further divided according to the profile of temperature change from month to month. A four month rapid heating phase occurs from September to December (August - September to November - December, Fig. 3.4) and is typified by a rate of temperature input at the surface in excess of the rate at which heat is transferred downward to the next sampled depth (6 m). In the remainder of the heating phase (January and February) the rate of surface heating is less than that at 6 - 18 m, indicating that the downward heat transfer exceeds the rate of surface heating.

Cooling begins in March and a zone of active heat loss gradually extends downwards between March and June, affecting the upper 10 m (approximately) at first and extending down to about 40 m in the May - June difference plot (Fig. 3.4). Over the same period there is a wave-like zone of heating below the cooling zone. This represents the continued downward propagation of the summer heat income, which appears to penetrate to 40 - 50 m before it is overtaken by the deepening zone of heat loss. The final phase is between June and August when active cooling affects the entire water column.

Fig. 3.2 indicates that there is slight heating of the deepest layer (nominally 82 m below the surface) which proceeds at an average rate of 0.10°C month^{-1} from January to April, but is faster (0.22°C month^{-1}) between September and December following the period of maximum vertical circulation; the rate from September to April is 0.13°C month^{-1}. In the period from June to September the deepest sampled layer cools at an average rate of 0.3°C month^{-1}. It should be noted that these long term means incorporate the effects of inflow.

STABILITY, WIND WORK AND HEAT CONTENT

Theoretical Considerations

Heat budget, stability and wind work are empirical quantities derived from thermal and salinity profiles, and a knowledge of lake morphometry. Appendix 1 contains information on the computational method used to calculate these quantities for the present study. Appendix 3 contains a submitted paper (concerning Deep Lake, Antarctica) which makes use of these concepts and from which some of the following information is taken. References which deal with the mathematical and conceptual framework of these quantities in more detail than is given in this text include Birge (1916), Hutchinson (1957), Idso (1973), Walker (1974), Johnson et al (1978), and Viner (1984).

Heat Budget:-

The Annual heat budget of a lake is the term used to denote the change in stored heat between the time of minimum heat content and the subsequent time of maximum heat content. It is the amount of heat stored by the lake during the annual heating period, expressed per unit surface area of the lake. In the southern hemisphere the sequential nature of Birge's heat budget often means that it is calculated as the difference between the winter minimum of one calendar year and the maximum occurring early in the following year. Stored heat is a function of mass, temperature and specific heat. In a freshwater system, mass and specific heat are unity, so that the heat content (calories) is calculated simply as temperature x volume (Wetzel and Likens 1979).

Wind Work and Stability:-

The concepts of wind work and stability, which have their basis in elementary physics, were formalised by Birge and Schmidt in the late 1800's and the early part of this century. Two quotes from Birge's (1916) paper on "The work of the wind in warming a lake" help to explain both the conceptual basis for the wind work, and the difference between this concept and that of Schmidt's stability of a lake.

"The work to be done in warming a stratum of water which lies below the direct influence of the sun is done against gravity which resists the descent of the warmer and lighter water. The net work done in warming a stratum to a given degree may be measured by the energy which would be needed to transport the mass of water, thus warmed, to the place where it is found, against the resistance of denser water .... We may think of such a stratum as pushed down to its place through water ... [denser water] ..., somewhat as a sheet of cork might be forced down to its place through the water."

"Schmidt's problem begins where the problem ends which is discussed in the present paper. He considers not the amount of work needed to produce the stability (and this is another way of stating my problem) but the amount necessary to continue the distribution of heat until an indifferent equilibrium again results." (This indifferent equilibrium refers to an isothermal water column).

Another way of viewing the two quantities is in terms of the "centre of gravity" of a lake. Considering the lake basin as a shape or container full of isothermal (iso-dense) material, air, water, or mercury, the centre of gravity of the container is defined simply by its shape (the distribution of volume with depth in the lake). If, however, the lake is density stratified (i.e. by temperature or salinity gradients) with relatively light material overlying denser material then the centre of gravity of this stratified lake lies lower in comparison with that of the unstratified lake. Birge essentially calculates the work done (against gravity) to lower the lake's centre of gravity, by pushing layers of warmer water down into the denser underlying layers, while Schmidt calculates the work required (also against gravity) to lift the mass of water in the lake back up to its original centre of gravity (see Birge 1916).

In summary, then, Birge's work of the wind and Schmidt's stability are complementary quantities expressing work done through the lake surface; firstly to establish a given density stratification from an assumed initial condition, and then to destroy that stratification without further exchange of heat or solute.

Birgean wind work is limited by its conceptual time dependence, in that a starting point is assumed, several months prior to the period of maximum summer stratification. Simplified assumptions are necessary concerning the processes which led to the final maximum departure from the initial condition. In the context of Lake Burragorang, the contribution of advection to this density stratification complicates the interpretation of the wind work term. In general, processes that contribute to density layering but do not operate through the surface of the lake (for example inflows of different temperature, turbidity or salinity), fall outside the conceptual framework of Birgean wind work.

Schmidt Stability, on the other hand, is conceptually instantaneous and is therefore not limited by assumptions concerning the establishment of the existing density stratifications. It does however assume that the means of destroying the stratification is turbulent energy transferred to the lake through its surface (ie by the wind). Finally, it should be noted that all these quantities are minima. That is, the Heat budget is the minimum quantity of heat stored in the lake during the period of heat income. Birgean Work and Schmidt Stability are minimum estimates of energy transferred to a lake through its surface by the wind (see Birge 1916; Idso 1973).

Introduction

Based on the monthly profiles depicted in Fig. 3.3, in conjunction with hypsometric data, whole-lake estimates of three thermal energy and mechanical stability parameters were made for Lake Burragorang: Heat content (cal cm^{-2}), Idso-modified Schmidt stability (gm-cm cm^{-2}), and Birgean wind work (gm-cm cm^{-2}). Table 3.1 contains annual quantities (annual heat budget, annual maximum stability and annual maximum wind work) for Lake Burragorang and some other of the world's lakes. The seasonal progression of these terms is shown for Lake Burragorang in Fig. 3.5.

Annual Heat Budget

The data presented in Table 3.1 covers a broad range both in terms of lake morphometry and of values reported for specific parameters. Lake heat budgets have been more frequently reported in the literature than either stability or Birgean wind work and there is, therefore, a greater and more organised data set with which to compare the results obtained for Lake Burragorang. Within the context of Table 3.1 Lake Burragorang has an annual heat budget which falls at the lower end of the range, with the exception of tropical Lakes Brokopondo and Valencia. Temperate lakes of comparable or lesser volume than Lake Burragorang, such as Lakes Washington and Mendota, have higher annual heat budgets. Even polar Deep Lake (hypersaline and ice-free) has a higher annual heat budget. ~~Gorcham~~ Gorham (1964) established relationships between several morphometric parameters and annual heat budgets, using data derived mainly from Hutchinson (1957). In relation to Gorham's (1964) regressions Lake Burragorang's heat budget is consistently low for a monomictic lake. Using mean depth, surface area, and lake volume as predictor variables Gorham's (1964) regressions predict an annual heat budget of 24000, 30000, and 29000 cal cm^{-2} yr^{-1} respectively. Lake Burragorang's heat budget, based on averaged annual budgets for the 20 individual years is 19000 cal cm^{-2} yr^{-1} and 16500 cal cm^{-2} yr^{-1} from the 20 year mean monthly temperature profiles. In the latter calculation the use of long term averaged monthly profiles smooths maximum and minimum

Table 3.1 Comparative data for annual heat budget, Birgean work, and mechanical stability, for a variety of lakes.

Lake	Latitude	Altitude (m. ASL)	Area (km^2)	Depth Maximum (m)	Depth Mean (m)	Annual Heat Budget (cal cm^{-2})	Birgean Wind Work (gm-cm cm^{-2})	Heating Efficiency (gm-cm cal^{-1})	Schmidt Stability (gm-cm cm^{-2})	Source
Burragorang	33° 55' S	117	75	105	27.4	16530	3080	0.186	3630	
Atitlan	14° 40' N	1555	136.9	341	183	22110	3741	0.169	21500	Hutchinson (1957)
Riñihue	39° 50' S	117	77.5	323	162	28412	6027	0.212	36941	Campos et al (1978)
Kamloops	50° 46' N	336	52.1	143	71	65000				Carmack et al (1979)
Pyramid	40° 10' N	1173	532	104	57	32925	3878	0.118	8948	Hutchinson (1957)
Powell	37° 30' N	1110	493	153	46.3	40000	8000	0.200	15000	Johnson & Merritt (1979)
Pang-gong Tso	33° 45' N	4241	279.2	50	26.1	22000	1320	0.06		Hutchinson (1957)
Deep (Ant.)	68° 34' S	-50	0.64	36	20.3	24630	9000	0.365	8100	Ferris & Burton (in press)
Valencia	10° N	420	355	39	19.0	4984	2782	0.558	370	Lewis (1984)
Washington	47° 40' N	6	128	65	18.0	43000				Hutchinson (1957)
Mendota	43° 07' N	259	39.2	25.6	12.1	24073	1209	0.066	514	Wetzel (1983)
Brokopondo	4° 26' N	37	850	35	10.6	3326	571	0.172	391	Heide (1982)
Lough Neagh	54° 35' N			34	8.6	13800				Gibson & Stewart (1973)

values of heat content, lessening the annual budget term by approximately 13% relative to the mean of annual heat budgets calculated for each of the 20 years individually. These budgets include the effect of changing lake volumes whereas the budget from the 20 year mean profiles does not. Although this affects the heat content, its effect on the annual heat budget (a relative term) is probably less than 5%.

Gorham (1964) also determined some effect, on the annual heat budget, of basin shape which he characterised by the ratio √area/mean depth. Lakes with a high ratio showed slightly lower heat budgets than those with low ratios. Lake Burragorang falls within the range of morphometries used by Gorham (1964) but resides amongst those with an unusually high ratio and may therefore be expected to have a somewhat lower heat budget than lakes of comparable volume but different morphometry. In contrast, Lake Powell, which also has a high ratio of area to depth, has a measured heat budget (Johnson and Merritt 1979) which consistently exceeds the budget predicted from its morphometry. This suggests that factors other than the area to depth ratio may be of over-riding significance in causing Lake Burragorang to occupy the lower boundary of the relationships between morphometry and heat budget.

Variation of the Annual Heat budget:-

Hutchinson (1957) observed that lakes under the influence of oceanic climatic regimes tended to have more variable annual heat budgets than lakes subject to a more continental climate. Timms (1975) found that the extent of year to year variation in heat budget was related to lake morphometry, such that variability increased with volume and mean depth, but was influenced more by mean depth than volume. He regarded the "passive components" (volume, mean depth, and area) as apparently more significant than climatic factors or the extent of protection from wind. Timms (1975) used the coefficient of variation (CV; Standard deviation expressed as a percentage of the mean) to compare the variability of the annual heat budgets of 23 lakes for which 3 or more years of data was available. Lake Burragorang, with a CV (for 20 years of data) of 13.9%, shows a greater variation than lakes of similar mean depth. One lake, Gmündener (Austria), of similar volume to Lake Burragorang but with a much greater mean depth, had a higher CV. Timms (1975) reported CV's ranging from 38.4% (Lake Zug, Switzerland) to 1.9% (West basin, Australia). Lake Burragorang's CV is greater than the median value (7.1%) for Timms' (1975) data (including Burragorang). Some of the variation in Lake Burragorang's heat budget is doubtless a product of advective processes. This is more fully discussed later.

Birgean Wind Work

Table 3.1 shows that the Birgean wind work for Lake Burragorang is comparable to Birgean wind work for Pyramid Lake and Lake Atitlan, both of which are much larger lakes. Smaller values of Birgean wind work characterise Lake Mendota, Pang-gong Tso, and tropical Lake Valencia. Lakes Riñihue and Powell have much larger Birgean wind work values than that found for Lake Burragorang. These values are larger than any reported by either Hutchinson (1957) or Wetzel (1983).

Heating Efficiency:-

Hutchinson (1957) used the ratio of Birgean wind work/summer heat income (equals annual heat budget in monomictic lakes: units of ratio, gm-cm cal^{-1}) as a measure of the efficiency of heating in lakes, noting that the ratio could be expected to be low (efficient heating) for polar lakes and large, deep, temperate lakes and small (inefficient heating) for tropical and small shallow lakes. The range of values reported in the literature is from 0.03 gm-cm cal^{-1} (Arctic, Lake Chandler; Hutchinson 1957) to 0.54 gm-cm cal^{-1} (Tropical Lake Valencia; Lewis 1984). The comparative inefficiency of heating in polar Deep Lake (Antarctica) is explained by its very great salinity (c. 10 x

Seawater), which increases the density change $^\circ C^{-1}$ requiring a greater expenditure of wind energy to mix the lake. The ratio for Lake Burragorang (0.186 gm-cm cal^{-1}: Table 3.1) lies within the range of reported values, though it exceeds all of the values reported by Hutchinson (1957). Lake Burragorang is comparatively inefficient, requiring a relatively large amount of work per calorie of heat absorbed by the lake. Table 3.1 indicates, however, that ratios as high or higher than that reported for Lake Burragorang are not unusual. If the tropical lakes are discounted, then Lakes Powell (0.2 gm-cm cal^{-1}) and Riñihue (0.212 gm-cm cal^{-1}) have ratios similar to that of Lake Burragorang; they also share some general morphometric features. Lake Powell is a reservoir occupying a flooded river valley (Johnson & Merritt 1979), and Lake Riñihue lies in an elongate morraine-dammed glacial valley (Campos et al 1978). It is possible that relatively deep and narrow lakes are characterised by inefficient heating compared to lakes where vertical wind-induced mixing is less impeded by contact with the lake sides. If this is the case then high Birgean wind work/summer heat income ratios may be anticipated for other man-made lakes.

Schmidt Stability

This section centres around data smoothed to single monthly profiles for the full 20 year study period and, as such, the maximum Idso-modified Schmidt Stability, reported in Table 3.1, is expected to be somewhat underestimated. Lake Burragorang is capable of stabilities greater than 4000 gm-cm cm^{-2} based on profiles from individual sample days. In Table 3.1 the reported value of 3626 gm-cm cm^{-2} is well within the range of values found from the limited number of years for which stabilities have been calculated using individual profiles.

Table 3.1 contains a broad range of Schmidt stability values but most exceed the value given for Lake Burragorang. A much smaller temperate lake

(L. Mendota) and tropical lakes Brokopondo and Valencia are less stable than Lake Burragorang. A possible reason for this is the volume distribution in Lake Burragorang. Less than 10% of the lake volume is found in the lower 60 m of the water column (ie below FSL minus 40 m). In the calculation of Schmidt stability, the contribution of each horizontal layer is weighted according to its volume so that despite their great distance from the lakes centre of gravity these deeper layers contribute little to the overall stability.

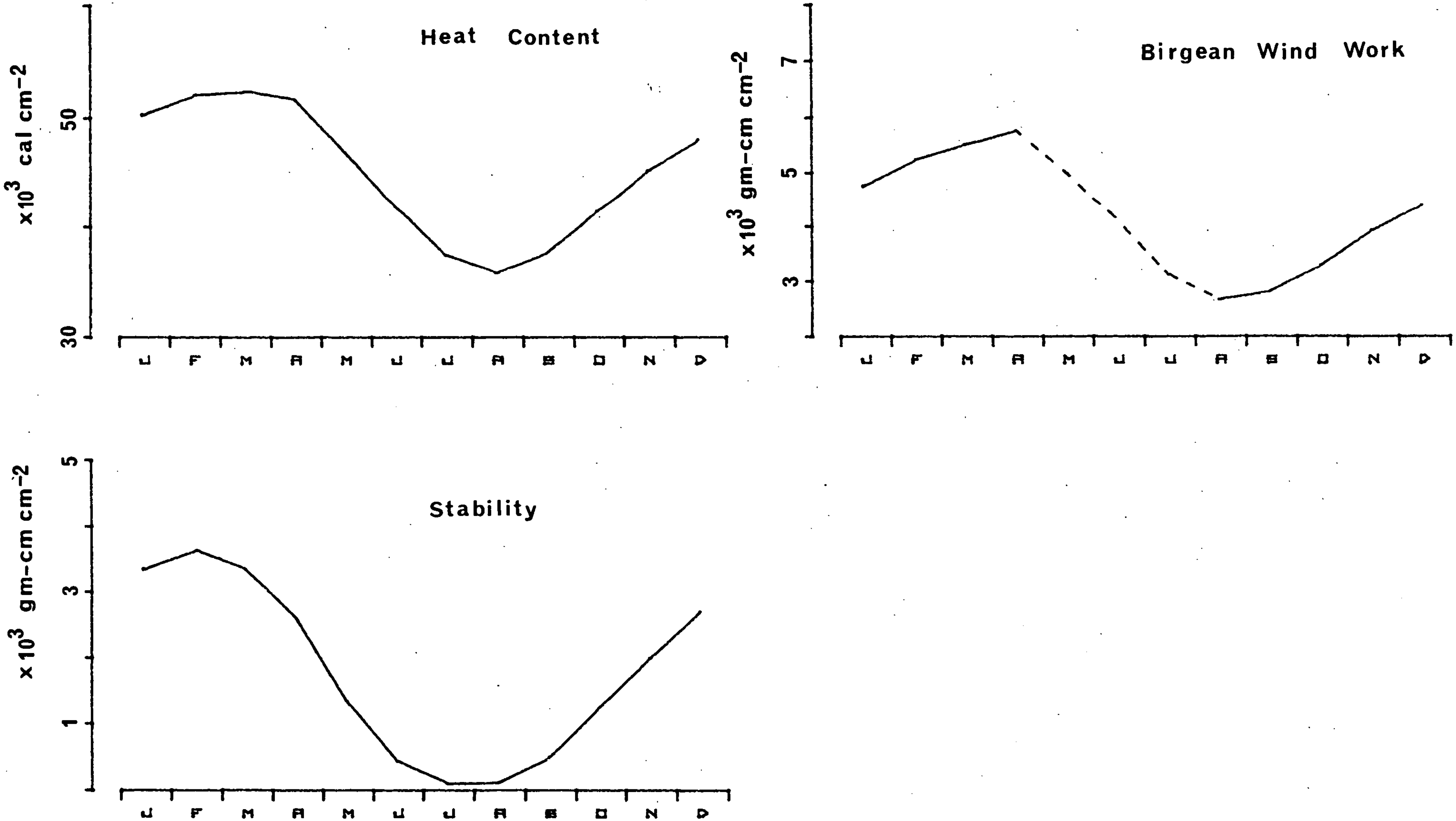
Viner (1984) found that the stability of lakes in New Zealand increased with mean depth, which is consistent with the information in Table 3.1 in that the more stable lakes (Atitlan, Riñihue, Powell, and Pyramid) all have considerably greater mean depth than Lake Burragorang (mean depth 27.4 m). Again, the exception is Deep Lake (mean depth 20.3 m) in which a relatively small thermal gradient can confer disproportionate stability because of the great density change $^{\circ}C^{-1}$. In fact, Deep Lake develops summer thermal gradients greater than many temperate lakes, which further enhances its stability.

<u>Seasonal Progression: Heat Content, Stability and Wind Work</u>

Fig. 3.5 shows that when the 20 year mean monthly temperature profiles for site 3D are used to estimate whole-lake heat content and mechanical stability a particularly smooth and orderly progression results. This is not the case when data from individual sample days is used. These show a comparatively "ragged progression" (Hutchinson 1957) and considerable variation is possible over periods of days. This may be especially true of Lake Burragorang because of the somewhat disproportionate contribution of the near surface layers to the total lake volume. Stewart (1973), in a detailed study of three Wisconsin Lakes (Mendota, Monona, and Waubesa), found that heat content in summer varied up to 10% within 2 - 3 days, representing gains and losses in excess of the entire winter heat budget.

FIGURE 3.5

Seasonal progression of heat content (cal cm^{-2}; x 4.19 = Joules cm^{-2}), Schmidt stability (gm-cm cm^{-2}; x 9.807 10^{-15} = Joules cm^{-2}), and Birgean wind work (gm-cm cm^{-2}), calculated using the profiles of monthly mean temperature (1961 - 1980; Fig. 3.3), and assuming the lake to be at full supply volume. The Birgean wind work is marked as a broken line during the cooling phase, in keeping with Birge's original concept of a number expressing the deviation of a lake's thermal stratification from a preceeding mixed condition.



Heat Content
x10^3 cal cm^-2
50
30
Birgean Wind Work
x10^3 gm-cm cm^-2
7
5
3
Stability
x10^3 gm-cm cm^-2
5
3
1
J F M A M J J A S O N D

Fig. 3.5 shows the three parameters peaking at different times in Lake Burragorang. The maximum Schmidt stability occurs in February, prior to that of heat content (March). That this should be so was predicted by Schmidt (1928) and was also suggested by Hutchinson (1957), but it is not always apparent for individual years, presumably because of the "noise" associated with short term variation of the three parameters. In their analysis of five North American lakes, Johnson et al (1978) found that the maximum stability preceded the maximum heat content for Lawrence Lake, but in Mirror and Findley Lakes the two maxima coincided (from monthly data for one year). The maximum Birgean wind work for Lake Burragorang is in April, following the maximum heat content (Fig. 3.5). In contrast, Johnson et al (1978) found the maximum Birgean wind work coincident with the maximum heat content for all of the lakes studied. From a four year study of Lake Powell, Johnson and Merritt (1979) present data indicating that the maximum Birgean wind work occurred after the month of maximum heat content in two of the four years and the two were coincident in the remaining years. Again however, the seasonal progression for single years is irregular when compared to long term averaged data. Minimum stability is recorded in Lake Burragorang in July, though it is only marginally less than the August stability. Both heat content and Birgean wind work are minimal in August.

Rates of Change

Fig. 3.5 shows that changes in the three parameters with time are nearly linear during heating and cooling, for periods of 3 - 4 months. Over the rapid heating phase (September - December) the heat content increases at an average rate of 113 cal cm^{-2} day^{-1} (assuming 1 month equivalent to 30.4 days). This is compared to heating rates from other lakes in Table 3.2. Lake Burragorang shows a heating rate comparable with the range of values reported for Windemere (118 and 200 cal cm^{-2} day^{-1}; the average is given in

Table 3.2 Heating and cooling rates for various lakes, (Change in heat content per unit time).

Lake	Heating Period	Heating Rate Mean (cal cm^{-2} day^{-1})	Heating Rate Maximum (cal cm^{-2} day^{-1})	Cooling Period	Cooling Rate Mean (cal cm^{-2} day^{-1})	Cooling Rate Maximum (cal cm^{-2} day^{-1})	Source
Burragorang (Aust.)	Sep. - Dec.	113	125	Apr. - Jul.	-157	-169	
Powell (USA)	Apr. - Aug.	368	439	Sep. - Jan.	-345	-392	Johnson & Merritt (1979)
Kamloops (Br. Col.)	June		c. 400	January		c. -400	Carmack et al (1979)
Baikal (USSR)	117 days	362					Hutchinson (1957)
Klammingen (Sweden)	May		332	October		-201	Hutchinson (1957)
Riñihue (S. Am.) [a]	December		300	July		-260	Campos et al (1978)
Green (USA)	May - Aug.	215					Hutchinson (1957)
Mono (USA)	May - Aug.	176					Mason (1967)
Ladoga (USSR)		160					Hutchinson (1957)
Windemere (Gr. Br.)		160					Hutchinson (1957)
Titicaca (S. Am.)	Sep. - Dec.	158		May - Jul.	-172		Richerson et al (1975)
Neagh (Ireland) [b]	May		106	November		-106	Gibson & Stewart (1973)
Augher (Ireland)	Apr. - May		76				Rippey (1983)
Deep (Antarctica)	Sep. - Dec.	252	444	Jan. - Sep.	-107	-339	Ferris & Burton (In press)

[a] These rates apply to the surface layer only.

[b] Predictions from an empirical model. content per unit time).

Table 3.2) and Lough Neagh. Higher rates are reported for Lakes Powell, Baikal, Green Lake, Deep, and Riñihue. A lower rate is reported for Augher Lough which is much smaller than any of the others. Over the same period (September - December) Schmidt stability increases at an average rate of 24 gm-cm cm^{-2} day^{-1} and Birgean wind work increases by 17 gm-cm cm^{-2} day^{-1}, though with only approximate linearity.

During the cooling phase the period of near linear decline is from April to July. Heat content declines at an average rate of -157 cal cm^{-2} day^{-1}, which exceeds the heating rate by 39% (difference as percent of heating rate). This is perhaps not surprising when it is considered that heating is a process largely confined to daylight hours, while cooling may be continuous. Rates of cooling are given for some of the lakes in Table 3.2. Gibson and Stewart (1973) used a sine-wave to approximate water temperature changes in Lough Neagh, and predicted equal maximum heating and cooling rates of 106 cal cm^{-2} day^{-1}. Campos et al (1978) reported a maximum surface cooling rate less than the maximum rate of heating in Lake Riñihue, and this is also evident for Lake Powell (data derived from graphs averaged over 4 years, from Johnson and Merritt 1979). Data taken from Richerson et al (1975) indicate a cooling rate in excess of the heating rate for Lake Titicaca. However, the difference between rates of heating and cooling for all these lakes is about 10% (of heating rate), markedly less than for the long term data from Lake Burragorang.

In this period (April - July) stability declines at a maximum rate of -41 gm-cm cm^{-2} day^{-1} between April and May, after which it declines more slowly (Fig. 3.5); between June and July the rate is -11 gm-cm cm^{-2} day^{-1}. That the wind work should decline simply because the lake is cooling is conceptually unsatisfactory, and is represented as a dashed line in Fig. 3.5, because Birge's original concept was of the work done by the wind in heating a lake and is not relevant to the cooling phase, although the direct work curves (depth profiles of Birgean wind work) are useful in this regard (see Johnson et al 1978; Johnson and Merritt 1979 and Johnson et al 1984).

VARIATION OF THE THERMAL CYCLE

In the previous section the thermal regime of Lake Burragorang was examined using the full data set condensed into 12 temperature profiles. The annual cycle of thermal stratification was discussed, as were derivative data concerning mechanical stability, heat budget, and the work of the wind. On this basis, Lake Burragorang was placed in the context of some other of the worlds lakes and reservoirs. Little attempt has so far been made to assess the variation in Lake Burragorang's thermal cycle or to describe the events that generate that variation. In this section the diversity of thermal behaviour, previously excluded by the long term averaging of thermal profiles, is examined.

The detailed variation of the thermal cycle, is apparent from Fig. 3.1. A very obvious feature of the temperature isopleth diagram is the irregular short term vertical movement of some isopleths in the region of the thermocline. Such movements are commonly associated with wind-forced harmonic motion of the thermocline, seiches (see Hutchinson 1957), which operate on a much smaller time-scale but are detected here from mainly weekly measurements. In man-made lakes these movements can also be caused by the opening and closing of the subsurface offtakes (Wunderlich 1971). Particularly distinct examples of these temporary displacements of the thermocline occur between October and December of some years, for example 1965, 1975, 1977, and 1978 when vertical movements are of the order of 5 - 10 m. The MWS&DB planned (c. 1978) to investigate seiches in Lake Burragorang, using fixed temperature recorders positioned at the lake sides (near site 3D), but the very low water levels that developed in 1979 and

1980 continued until about 1984 so that data collection has only recently become possible.

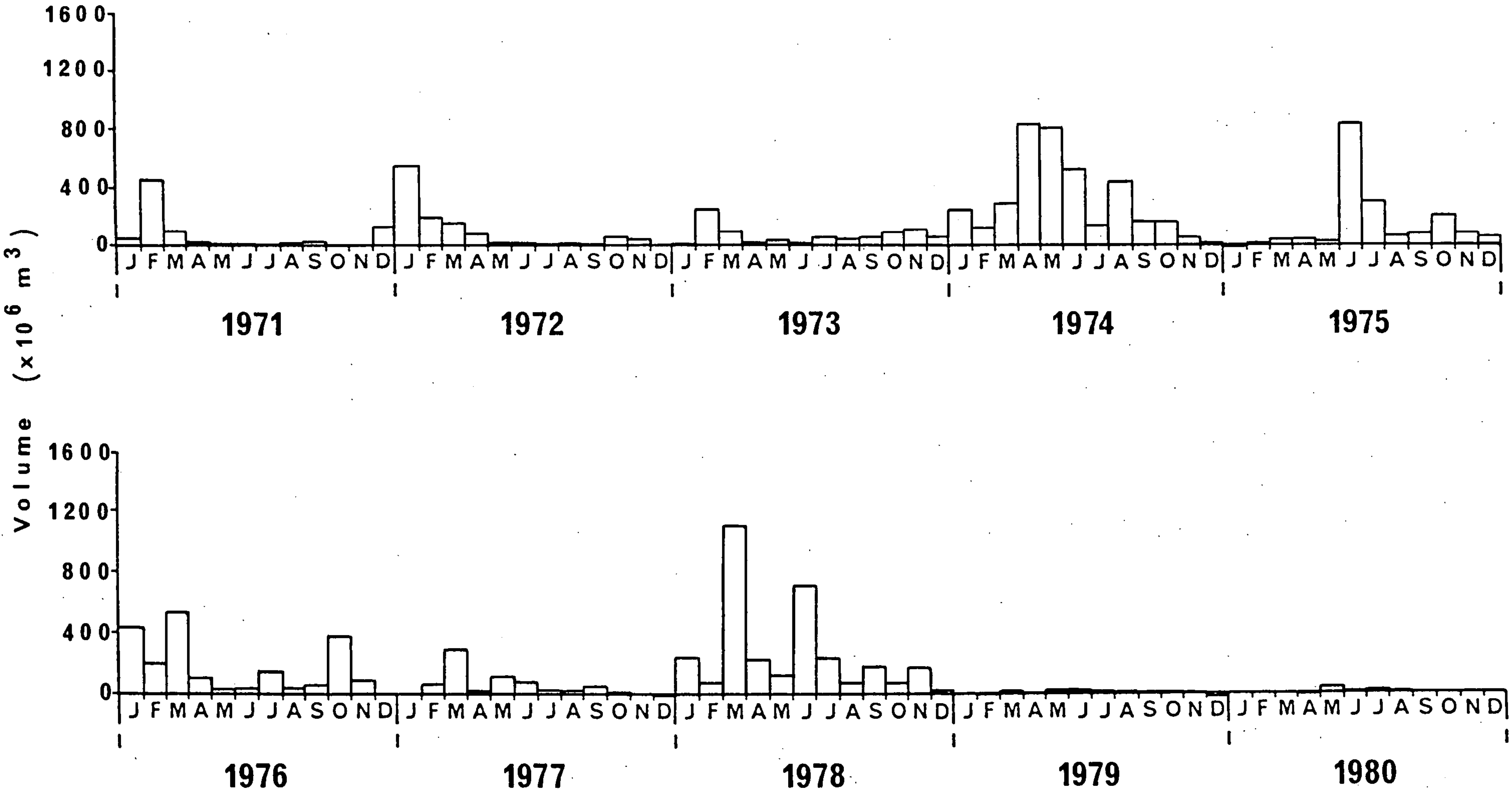
Quite significant differences in the slope of the summer metalimnial gradients are apparent from Fig. 3.1. Tightly bunched isopleths, indicating a steep thermal gradient, are characteristic of 1965 and 1968, whereas 1963 and 1972 show a much broader spacing of isopleths (in summer) indicative of much gentler thermal gradients. The depth of the epilimnion and its annual progression are similarly variable from year to year. In 1965 and 1968 the epilimnion remains more or less stable from the time of its formation in the previous year to late autumn. This is indicated by isopleths approximately parallel to the water surface (Fig. 3.1). In comparison the 1963 - 1964 and 1976 - 1977 stratification periods show a nearly continuous deepening of the epilimnion between September, when the lake re-stratifies, and the following May. A third variation is exemplified in 1975 - 1976 and 1977 - 1978, when there are step-like increases in the depth of the epilimnion during the summer and autumn.

A particularly important observation from Fig. 3.1 is the existence of cooling periods, in deep water, which are independent of surface temperature changes. These events are most readily recognised from the appearance of the 11°C isotherm below about 40 m in the winter to spring of 1961, 1963, 1964, 1967, 1975, and 1978. Any temperature change in deep water would, in a purely convective system, be mediated by changes in the temperature of the overlying water. Consequently, times when the deep layers are cooled independently of the temperature of the overlying water are indicative of advective effects. For the study period, water less than 11°C in the deep layers at site 3D has always been the product of cold underflows. Depending on the ambient hypolimnetic temperature, the 12°C isotherm occasionally has similar significance.

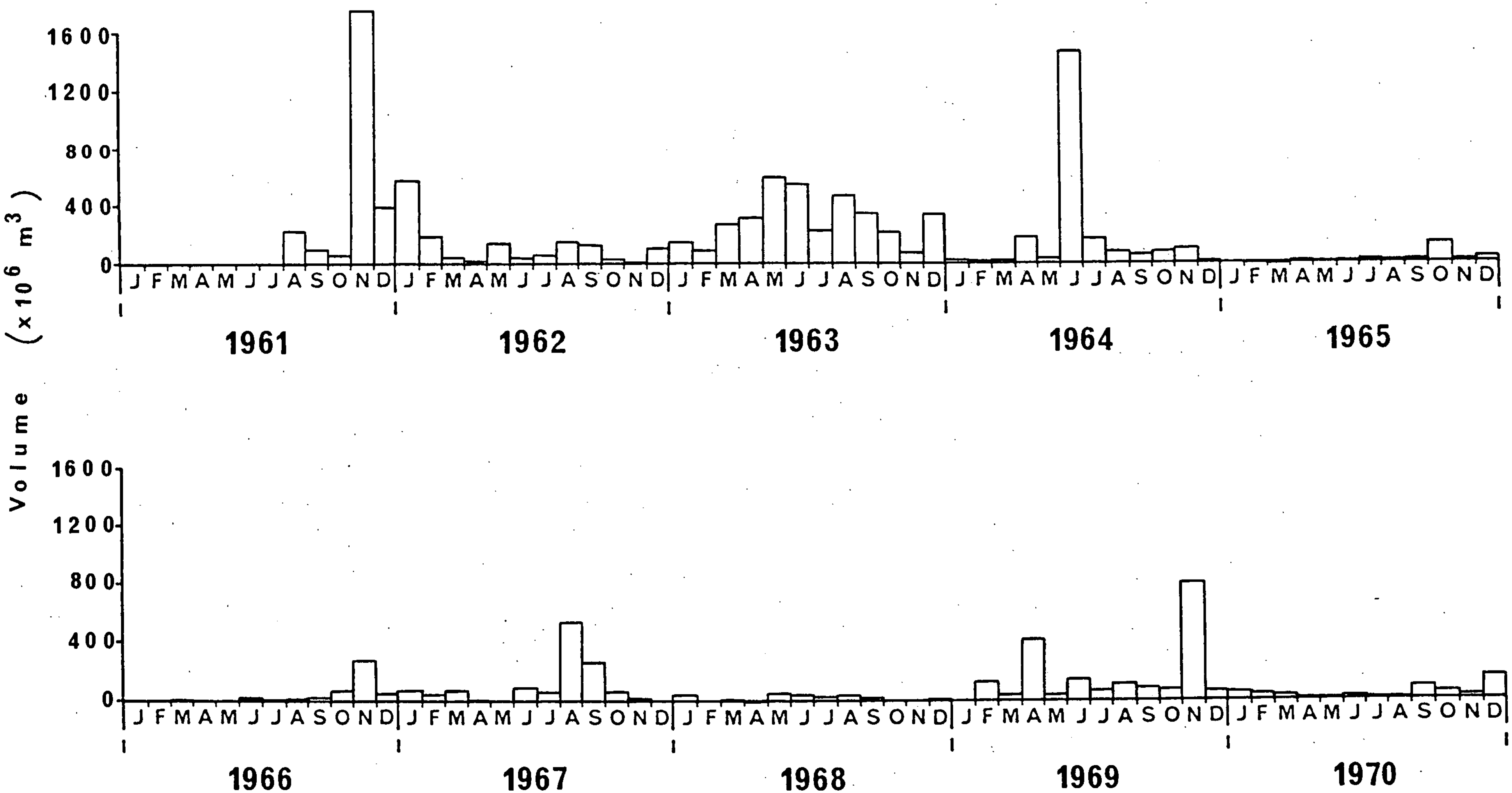
In Fig. 3.6 monthly total inflow (minus evaporation), derived from records

FIGURE 3.6

Monthly total inflow volume x 10^6 m^3 (minus evaporation) for the period August 1961 - December 1980. The time axis has the same scale (as nearly as possible) to that of the isopleth diagrams, to facilitate comparisons. The inflow (minus evaporation) has been calculated from records of outflow, taking into account changes in the reservoir level.



Volume (x10^6 m^3)
1600
1200
800
400
0
J F M A M J J A S O N D
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980


Volume (x 10^6 m^3)
1600
1200
800
400
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

of outflow and reservoir level (see Methods, Chapter 2), is plotted on a scale as close as possible to the time-scale of the isopleth diagram (Fig. 3.1). Direct comparison of Fig. 3.1 and Fig. 3.6 clearly demonstrates the association between the localised deep water cooling (ie the 11°C isotherm) and inflows. Particularly good examples are seen in August 1967 and July 1975. Detailed examination of the two figures indicates that the step-like deepening of the thermocline is also associated with inflow, for example March - April 1976 and March - April 1978. Clearly however, the zone affected by these autumn inflows is situated in the metalimnion rather than the hypolimnion as is the case for winter inflows.

An unusual and profound advective temperature effect occurred in 1961, in response to the largest flood recorded for the 20 yr study period. The inflow of November 1961, which first filled the lake, affected the whole water column at site 3D raising the hypolimnetic temperature to greater than 15°C (Fig. 3.1). No subsequent inflow has generated such an effect at site 3D.

It is apparent from these observations that the effect of inflow on the thermal structure of lake Burragorang should be examined in more detail.

INFLOWS AND THE CYCLE OF THERMAL STRATIFICATION

Inflow into Stratified Lakes

The observation that inflows into lakes may preserve a distinct identity, rather than mixing vertically with the lake water, is attributed to Forel (1892; reported in Wright and Nydegger 1980), in his classic study of Lac Leman, in the latter part of the nineteenth century. More recently, the study of density currents in large reservoirs has shown that inflows may traverse the full length of large lakes and still retain many physical and chemical characteristics (ie dissolved oxygen, turbidity, alkalinity, pH, and nitrate nitrogen; Weibe 1939) that distinguish them from the lake water. The long-standing interest in these currents, in reservoirs, arises primarily from their effects on water quality and sedimentation.

Wunderlich (1971) states the reason for the existence of discrete inflows as follows, "In the presence of density gradients, vertical movements of water particles are inhibited or suppressed while horizontal movements are enhanced. Consequently, while vertical mixing and dilution become small, horizontal movements become more pronounced and persistent. For this reason, inflows entering stratified reservoirs tend to retain their identity to a much greater extent due to reduced mixing than they would in homogeneous water bodies. Chemical and biological processess active in the inflow water also may remain limited to certain water parcels or layers. Also, outflows are withdrawn from zones at elevations corresponding approximately to the intake openings."

Assuming that temperature is the main determiner of density in both the lake and the inflow and that they do not have exactly the same temperature, there are two basic positions available to an inflow entering an unstratified lake. If the inflow is warmer than the lake water it will overflow, and if the inflow is colder it will underflow the lake water. If however the lake is thermally stratified, as is the case in Lake Burragorang for most of the year, then the inflow may occupy a third position in the metalimnion (interflow) if its temperature lies between that of the surface and deepest water in the lake (see Wunderlich 1971; Wunderlich and Elder 1973).

<u>Introduction</u>

In this section the effect of inflows on the warm monomictic thermal cycle of Lake Burragorang is considered, and also the seasonal interchange in the type of inflow (ie interflow, underflow) that affects site 3D. An important part of this investigation was the establishment of a register of "effective" inflows. This was based primarily on the profiles of the various water quality parameters, particularly turbidity and chloride, routinely determined at site 3D. Those inflows that affected water quality (including temperature and dissolved oxygen also) at site 3D and were detected as individual events are recorded in Table 3.3. The somewhat subjective criteria that were used to distinguish these inflows are enlarged upon in the Materials and Methods (Chapter 2).

Advective Stabilisation

The term "advective" is used to include the effects of both inflow and outflow, which may be regarded as essentially horizontal processes as opposed to the vertical processes of convection. The present discussion focusses on inflow, leaving consideration of outflow for a later chapter when data is presented which more clearly indicates the role of outflow at site 3D. It should be noted, however, that outflow may make a significant contribution to the events described below and that the two processes are interdependent, especially as the largest capacity offtake (the hydro-electric offtake; HEPS) is operated only when inflows raise the water level to a point where its use does not threaten reserves of potable water.

This effect of inflows on the annual cycle of thermal stratification is most clearly observed from the data for 1967. In the first half of the year monthly inflows were comparatively small, and thermal stratification proceeded in a manner typical of a warm monomictic lake. The depth of the epilimnion remained less than 18 m until mid-June, when a metalimnetic temperature gradient of 3.2°C (18 - 49 m) separated the epilimnion and hypolimnion. On the 1/8/67 overturn had still not occurred, although the epilimnion extended to about 40 m and a metalimnetic temperature gradient of < 1°C separated the epilimnion (c. 80% saturated with oxygen) from the hypolimnion which was from 5 - 12% saturated with oxygen.

The record of monthly inflows (Fig. 3.6) shows that a considerable inflow

Table 3.3 Inflow Register, 1961 - 1980. A list of inflows that were detectable by their individual effect on the water quality, at site 3D. The procedure used to select these inflows is described in the Methods, Chapter 2.

Inflow Initiation Date	Date of Maximum Recorded Effect Turbidity	Date of Maximum Recorded Effect Chloride	Total Inflow (minus evaporation)[a] $\times 10^6 m^3$	Inflow type	Relative Effect Turb.	Relative Effect Cl^-	Maximum Effect Turbidity Hellige units	Maximum Effect Chlor. $mg\ l^{-1}$
3/61	4/4/61	11/4/61		U	+	+	15	25
7/61	8/8/61	22-30/8/61		U	+	+	25	34
25/8/61	5/9/61	12/9/61	198.4	U	+	-	11	24
16/11/61	24/11/61	30/11/61	1582.7	U	+	-	700	7
12/7/62	26/7/62	26/7/62	40.8	U	-	+	9	17
14/8/62	4/9/62	11/9/62	131.5	U	-	+	7	24
29/4/63	14/5/63	28/5/63	597.1	I	+	-	40	11
5/6/63	18/6/63		446.5	I	+		30	18[b]
26/6/63	9/7/63	2/7/63	94.3	U	-	-	9	11
12/7/63	23/7/63	23/7/63	123.8	U	+	+	20	32
27/8/63	10/9/63	10/9/63	564.0	U	+	-	140	16
9/12/63	17/12/63		264.4	I	+		9	18[b]
10/6/64	16/6/64	16/6/64	1346.2	U	+	-	220	6
16/7/64	11/8/64		128.0	U	-		11	14[b]
8/11/66	22/11/66	15/11/66	261.9	I	+	+	4	25
6/8/67	22/8/67	15/8/67	493.4	U	+	-	250	9
15/4/69	28/4/69		340.8	I	+		9	25[b]
13/11/69	24/11/69	1/12/69	545.0	I	+	-	8	17
1/9/70	22/9/70		57.9	U	+		10	13[b]
9/2/71	15/2/71	15/2/71	173.2	I	+	-	13	16
14/1/72	7/2/72	6/3/72	552.2	I	+	-	8	13
4/3/72	20/3/72		141.9	I	+		15	20[b]
26/2/73	5/3/73	5/3/73	112.2	I	+	+	6	24
11/7/73	23/7/73	6/8/73	35.0	U	+	-	3	13
11/3/74	18.3.74		188.9	I	+	-	8	15[b]
20/4/74	13/5/74	13/5/74	589.7	I	+	-	15	14
26/5/74	5/6/74	18/6/74	812.1	U	+	+	45	20
27/8/74	9/9/74		515.7	U	+	-	80	17
21/6/75	30/6/75	30/6/75	834.1	U	+	-	200	11
15/7/75	4/8/75	28/7/75	125.5	U	+	+	16	25
23/2/76	8/3/76	8/3/76	466.3	I	+	-	13	11
17/10/76	1/11/76	22/11/76	356.7	U	+	+	40	23
28/2/77	7/3/77	7/3/77	216.4	I	+	-	12	14
18/3/78	28/3/78	3/4/78	1083.0	I	+	-	100	14
10/4/78	17/4/78	24/4/78	119.6	I	+	-	100	14
1/6/78	19/6/78	19/6/78	562.8	I	+	+	70	12
20/6/78		17/7/78	323.5	U		+	20[b]	23

[a] From the initiation of the inflow to the date of maximum recorded turbidity.
[b] No obvious change in this parameter, so the ambient value at the depth and time of the maximum effect for the other parameter is reported.

occurred in August 1967, and this is recorded in Table 3.3 as a single event initiated on the 6th of August and penetrating to site 3D as a turbid underflow by the 8th of August. Apart from its turbidity this underflow was also some 2°C colder than the hypolimnetic water at site 3D, and consequently significantly denser. In Fig. 3.7 a series of profiles (Temperature, D.O., turbidity, and chloride) show that complete vertical mixing did not occur. The effect of inflow (and outflow) was to maintain density stratification of the water column for the brief period in which overturn may otherwise have occurred; this advective stabilisation persisted long enough for surface heating to again restrict vertical circulation to the near surface layers.

Detailed examination of all the available physico-chemical profiles strongly suggests that advective stabilisation has prevented complete vertical circulation (at site 3D) in 10 of the 20 years from 1961 - 1980. Comparison of the approximate periods of complete vertical circulation (or periods when the circulation extended to the deepest sampled depth at site 3D; marked on Fig. 3.1) with the monthly inflow totals (Fig. 3.6) indicates that inflows (and therefore outflows) during or just prior to the period of potential overturn (July - August approximately) are frequently effective in restricting vertical circulation long enough for re-formation of the near surface thermocline. It should be noted that even in years of advective stabilisation the volume of water excluded from circulation is always less than 10% of the total lake volume, and that the isolated water is invariably the most recent water to enter the lake. This second point means that the old hypolimnetic water is always displaced upward and mixed into the overlying, circulating water column. Advective stabilisation has occurred when the volume of the specific underflow has been relatively small, for example in July 1973, an underflow of 35 10^6 m^3 was associated with incomplete vertical mixing.

FIGURE 3.7

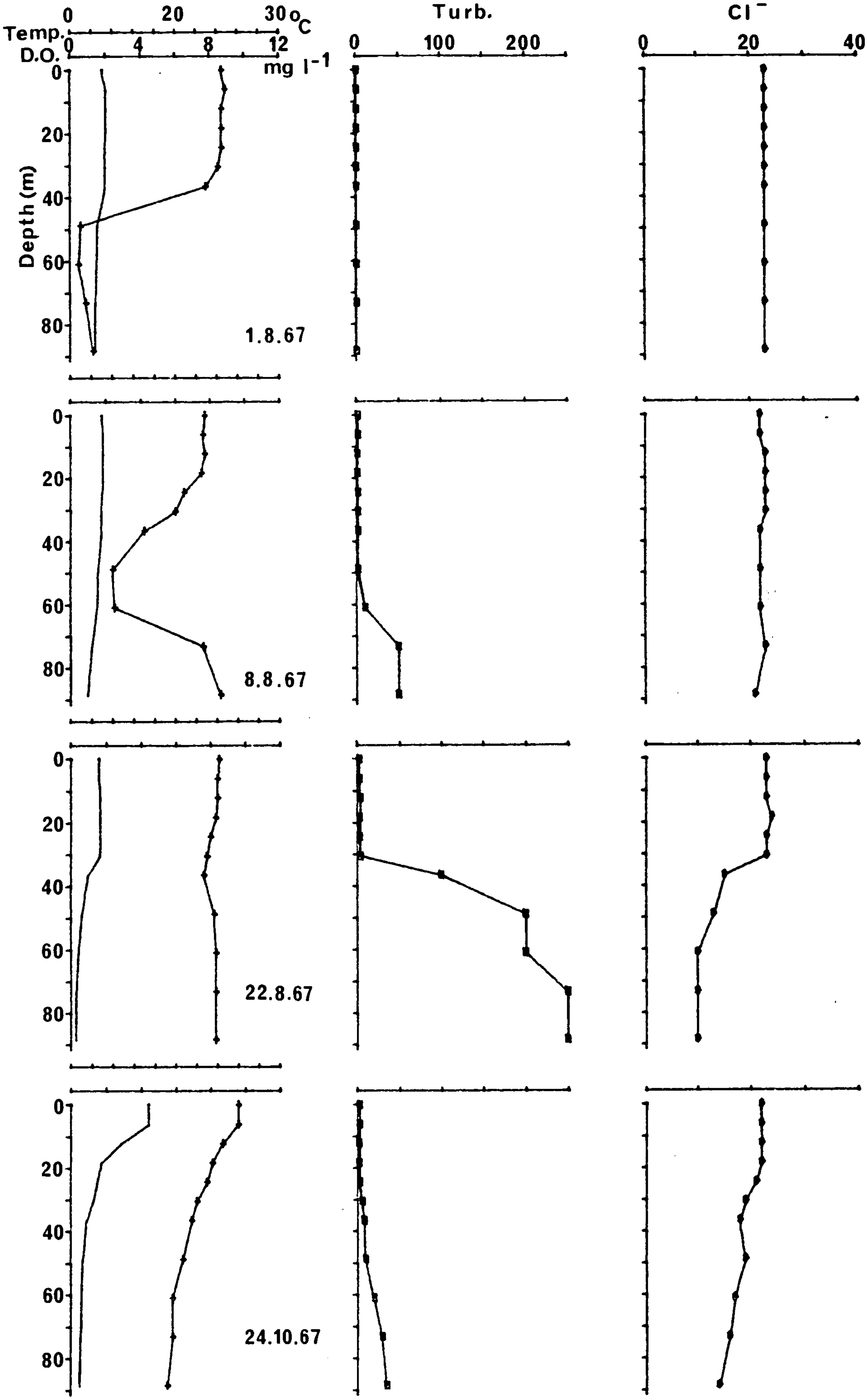
Profiles of temperatures (^{0}C) and dissolved oxygen (mg l^{-1}), turbidity (Hellige units), and chloride (mg l^{-1}), showing the effects of a cold underflow initiated on 6/8/67. Each row of three plots shows data from one sample day. The profiles for the 1/8/67 show conditions prior to the flood. Increasing turbidity is evident on the 8/8/67, as is the increase in oxygen concentration as the underflow displaces the old hypolimnetic water upwards. On the 22/8/67 all the parameters are affected, with hypolimnetic cooling and re-oxygenation apparent, as well as high turbidity and reduced chloride concentration.

Symbols:-

Temperature (·)

Dissolved oxygen (+)

Turbidity and chloride are on separate plots.



Temp.
D.O.
0 20 30 °C
0 4 8 12
mg l⁻¹
Turb.
0 100 200
Cl⁻
0 20 40
Depth (m)
0 20 40 60 80
1.8.67
8.8.67
22.8.67
24.10.67

There are some anomalous years which do not fit neatly into the scheme given so far. The data for 1969 is unfortunately variable, this is evident for the thermal data in Fig. 3.1, but also applies to other physico-chemical profiles making it difficult to reliably assess whether overturn occurred or not. Inflows did occur in June and August of 1969, so it is possible that 1969 was characterised by advective stabilisation, but this cannot be verified. In 1979 consistently low inflow is combined with a very short or incomplete overturn. In either case it may be reasonable to conclude that complete vertical mixing (at site 3D) is a tenuous event in some years and that this deep, narrow and elongate arm of the reservoir, which has valley sides rising 100 - 300 m above the water level, is morphometrically predisposed to being poorly wind-mixed in its deepest layers (the lower 40 - 60 m of the water column).

Advective Stabilisation in Other Lakes

Having established that advective stabilisation is a real and recurrent, if morphometrically assisted, event in Lake Burragorang, it is necessary to consider other examples of the same or similar behaviour in the World's lakes.

The clearest recognition of the potentially stabilising effect of inflows comes from research into stratification in tropical systems. Beauchamp (1969) suggested that cold inflows play a part in the apparently permanent stratification of Lake Tanganyika, and possibly in the initiation of stratification in Lake Victoria. Fish (1956; reported in Talling 1963) suggested this mechanism for thermal stratification in Lake Albert, but Talling (1963) considered this unlikely preferring a convective origin caused by profile-bound density currents. Gliwicz (1976) demonstrated advective stabilisation for a tropical impoundment with a short residence time, concluding that stratification of "kinetic origin" is a particularly important mechanism in the tropical zone, not simply as a stratification steepening agent, but also as an initiator of thermal stratification in tropical lakes and impoundments. Neel (1963) inferred the occurrence of advective stabilisation in temperate, man-made lakes of the middle United States; "Some studies of sizeable impoundments, however, indicated that significant homogenization of distinctive inflows may not be relied upon even when stratification is absent or ephemeral ...", based on reports from the early to mid 1950's. Larson (1979) details the occurrence of meromictic conditions (in an Oregon reservoir) which were initiated by a large turbid inflow and persisted for about 8 years. Steane and Tyler (1982) observed advective stabilisation (attributed to a cold underflow) in Lake Gordon (Tasmania), in 1978, and considered that it was likely to be a recurrent phenomenon. It would appear that the prevention of complete vertical mixing by cold underflows has yet to be reported for a natural temperate zone lake. Irwin and Pickrill (1982) and Pickrill and Irwin (1982) describe complex patterns of inflows into Lakes Tekapo and Wakatipu (New Zealand) but in both cases complete mixing occurs annually.

Several factors contribute to the repeated occurrence of advective stabilisation in Lake Burragorang.

1. The temperature of inflowing water invariably falls below that of the deepest strata in the lake during winter.
2. Large inflows can occur at the critical time prior to the normal period of maximum circulation in Lake Burragorang.
3. The depth and morphometry, particularly at site 3D, combine to ensure the stability of weak, but deep, density gradients.
4. Following from the previous point is the fact that the lake is characterised by a very short period when complete vertical mixing is possible.
5. As previously mentioned the general coincidence of inflows and outflow from the HEPS offtake may be significant (cf Johnson and

Merritt 1979).

These factors are also generally true for Lake Gordon, the other Australian impoundment for which underflows are reported to prevent holomixis (Steane and Tyler 1982). The other feature common to these lakes is that they have warm monomictic thermal cycles, and this is almost certainly true of the impoundments referred to by Neel (1963) and also seems likely for Hill's Creek reservoir (cf. Larson 1979). Killworth and Carmack (1979) and Carmack et al (1979) detail the lake/river interactions in dimictic Kamloops Lake (British Columbia), and despite the complex patterns of density flows the annual dimictic circulation was apparently unchanged. This greater complexity results from the fact that both the river and lake waters pass through the temperature of maximum density (4°C) twice annually, and interestingly this means that influent water will tend to overflow in winter, opposite to the anticipated bahaviour for most warm monomictic lakes.

A factor which is difficult to adequately evaluate is the role of morphometric pre-disposition to being poorly wind-mixed. The effect of limited wind access to the surface of a lake, as a result of its small surface area relative to its maximium depth or because of the surrounding topography or vegetation, is shown by the existence of biogenic meromictic lakes with vanishingly small meromictic (chemical) stabilities (< 1 gm-cm cm^{-2}; Hongve 1980). Similarly, Salonen et al (1984) draw attention to the importance of relative depth, humic content, orientation to prevailing winds, and the protection afforded by trees, to explain the frequent occurrence of meromixis and spring-meromixis in small forest lakes of Southern Finland. Steane and Tyler (1982) considered that the sheltered nature of Lake Gordon's western end would restrict the depth of wind-mixing to about 70 m. In oceanic systems the winter mixed depth may be taken to indicate the outer limit of wind generated vertical mixing in an essentially unconfined basin, although saline gradients may play a part in restricting vertical mixing. Denman and

Gargett (1983), in a theoretical analysis of the scale of large eddies (of comparable size to the vertical extent of the mixed layer), calculated scales of c. 40 m under non-stratified conditions. On the other hand Venrick (1984) reported finding a water mass, in the North Pacific, with an isothermal layer 140 m deep, while Heath (1984) reported that the winter mixed layer off New Zealand sometimes occupied the upper few hundred metres. In the confines of lake basins the depth to which wind-induced mixing can extend is well illustrated by the studies of the very deep oligomictic or meromictic lakes of Africa, Europe, and North America. Talling (1969) reported that mixing was confined to the the upper 200 m (approximately) of lakes Tanganyika and Malawi, despite a vertical temperature gradient of only 0.5°C in Lake Malawi. In Europe, several deep oligomictic lakes are known in which complete vertical mixing occurs only when particularly cold conditions and strong winds combine. Tonolli (1969) found that Lake Maggiore (maximum depth, 370 m) circulated fully in the winter of 1962 - 1963, but that mixing was generally restricted to the upper 150 - 200 m in other years. Similar oligomictic behaviour was also noted for lakes Como, Garda, Ohrid, and possibly Geneva (Tonolli 1969; Cole 1975). Weiss et al (1979) reported that Lake Constance was mixed to less than 200 m in the winter of 1963 - 1964, and noted that the water below 100 m was almost unaffected by seasonal heating. In North America, Wetzel (1983) observed that Great Bear Lake (Canadian Northwest Territories) only mixed below c. 200 m in cold years, when it circulated to its maximum depth of 450 m.

In general then, the wind is not unlimited in its ability to mix lakes vertically, even under conditions of very slight apparent stratification. However, a wide range of values is reported for the maximum mixed depth of deep lakes (non-meromictic); anywhere from less than 100 m to 200 m appears to be common, with mixing extending to much deeper layers in years with the appropriate weather conditions.

It has already been suggested that Lake Burragorang (at site 3D) may be so deep and narrow, and oriented such that full winter circulation will not occur in every year, regardless of advective stabilisation, it is therefore possible that underflows are simply features by which this morphometric predisposition to being poorly wind mixed is recognisable, rather than being significant in the prevention of holomixis. This question is implicit in Steane and Tyler's (1982) discussion of stabilisation by a cold underflow in Lake Gordon, and they apparently believe that the underflow of 1978 was not active in the prevention of holomixis in that year. Although this question cannot be definitely answered without more detailed observation of mixing processes in Lake Burragorang, there is sufficient evidence to assert that the lake is capable of complete vertical mixing (see Fig. 3.1 1966), and it is also apparent that winter underflows commonly cause the formation of deep density gradients. Further it is evident that very small density (thermal) gradients can act as effective barriers to vertical mixing in the region below about 50 m. Altogether, it seems likely that underflows actively promote the stabilisation in Lake Burragorang, but that morphometry is also a contributing factor. Again, the potential effect of currents generated by outflow must also be bourne in mind.

SEASONAL PATTERN OF INFLOWS

Inflows to Lake Burragorang, that penetrate to site 3D, are not always underflows. A seasonal cycle is evident, with regular and predictable interchange between interflows and underflows; overflows have not been observed at this site, presumably as a result of their entry into the actively mixing epilimnion. This would tend to restrict their influence to regions of the lake near the main inflows (cf. Ruttner 1963), although overflows have been observed to penetrate for a considerable distance from the point of inflow in some lakes (see Johnson and Merritt 1979; Hoffman and Jonez 1973). The seasonal distribution of interflows and underflows is shown, by the Inflow Register (Table 3.3). Underflows are generally a feature of the late autumn and winter months (c. May - September), although they have occasionally been recorded at other times, eg. in November 1961. Some predictive understanding of the seasonal occurrence of inflows is possible if it can be assumed that temperature is the dominant factor controlling the density of the inflowing water, and therefore the level at which the flow penetrates the water column. Both the sediment load and the concentration of dissolved material will also affect the density of the inflowing water. However, the contribution of dissolved material is probably slight, as the range of chloride concentrations in Lake Burragorang is small. A more significant contribution may be expected from the suspended material usually carried by inflowing water. The possible magnitude of this contribution can be inferred from the fact that a small inverse thermal gradient developed at site 3D in the winter of 1962, following the floods of November 1961. An inverse gradient of c. 1°C was associated with a turbidicline of from 300 to 400 Hellige units. This probably represents a maximum contribution, as these high turbidities have not been recorded in Lake Burragorang since that time (Table 3.3). In other studies of lake/river interactions Carmack et al (1979) concluded that temperature was the most important factor determining water density in Kamloops Lake (British Columbia), and Pickrill and Irwin (1982) drew the same conclusion for Lake Wakatipu (New Zealand).

Prediction of Interflows and Underflows

Assuming, then, that temperature controls the density of water flowing into Lake Burragorang, the long term mean temperatures of the two main inflowing rivers, compared to the surface and bottom temperatures recorded at site 3D, provides a simple model for the prediction of the seasonal progression from interflows to underflows and back again which is apparent from the Inflow Register (Table 3.3). In Fig. 3.8, the cycle of monthly average temperature in the Wollondilly and Cox rivers, are superimposed on the surface and deep-water temperatures recorded at site 3D, for the period 1961 - 1979. The figure shows that the Cox River is, on average, 0.5 - 1.0°C cooler than the Wollondilly River, and that the difference between mean and minimum inflow temperatures, for each river, is least in winter (c. 2 - 4°C) and largest in summer (c. 6°C). Pickrill and Irwin (1982) also noted a greater temperature range in summer, from continuous records for the Rees River, which flows into Lake Wakatipu (New Zealand). Based on the monthly averages, underflows are expected to occur from June to August (inclusive), when the temperature of both major inflows falls below that of the coldest water in the lake. However, it is a reasonable assumption that periods of high inflow will be characterised by temperatures less than the overall monthly means, particularly in summer. Pickrill and Irwin (1982) state "In winter the flood flows are commonly at about mean river temperature for the season ..., whereas in spring/summer they are commonly at or below the normal diurnal temperature ...". When the absolute monthly minimum temperatures, for the two inflows, are taken into account, the underflow period is extended to include May and September, and possibly April for the cooler Cox River inflow. If the potential contribution of turbidity (equivalent to about 1°C) is also considered, then it is conceivable, though not likely, that underflows could occur in both April and October, making a total underflow period of 7 months.

The broad predictive capability of this simple model can be tested by direct comparison with the Inflow Register. Figure 3.9 shows the distribution of the 37 effective inflows recorded at site 3D for the study period, based on the date of maximum recorded turbidity at site 3D (and not on the inflow initiation date, which occurs about two weeks earlier, on average). Of these, 20 were underflows, and 17 interflows (see Table 3.3). The distributions of

FIGURE 3.8

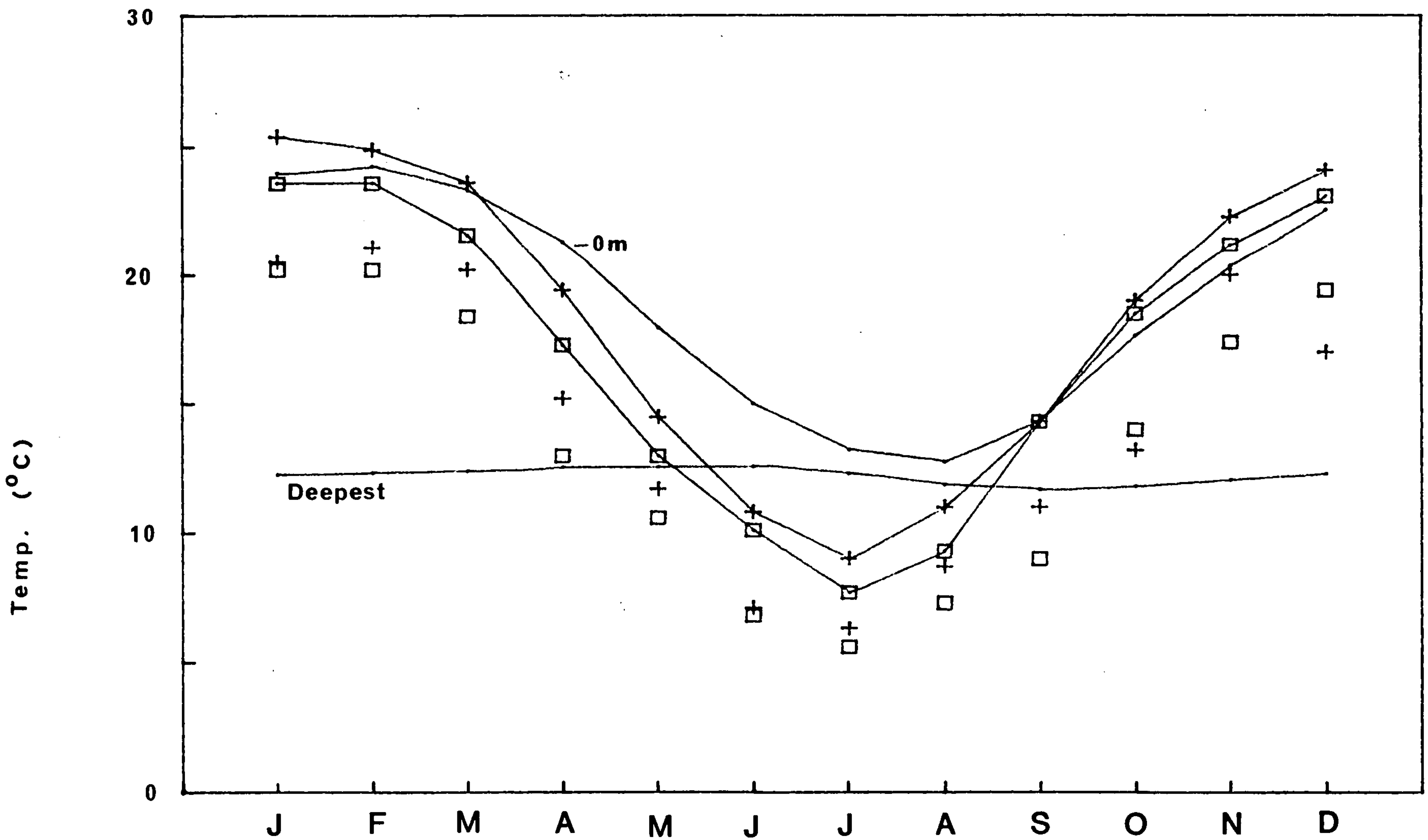
The temperatures (^{0}C) of the Cox and Wollondilly Rivers, measured at the inflow sample sites (see Chapter 2, Fig. 2.1) c. 50 km upstream from site 3D, plotted over the surface and deepest-sample temperatures at site 3D. All temperatures are monthly means for the period 1961 - 1980. Symbols not connected by lines depict minimum temperatures for each month (ie minimum of the c. 20 individual means for each month). The two rivers differ somewhat in temperature, with the Cox River generally cooler for most of the year.

Symbols:-

Wollondilly River (□)

Cox River (+)

Site 3D (·)

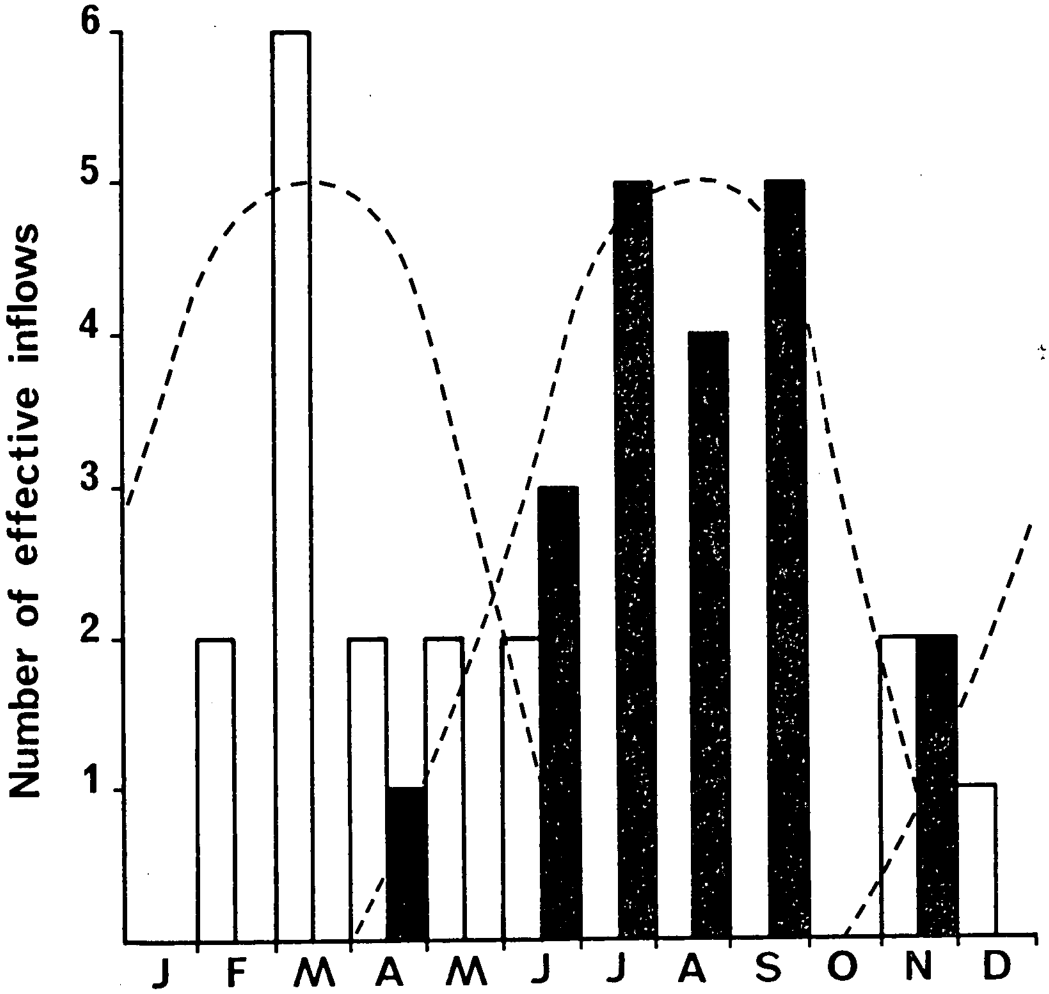


Temp. (°C)
30
20
10
0
0m
Deepest
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 3.9

Frequency histogram of registered inflows (see Table 3.3), showing the seasonal distribution of 17 interflows and 20 underflows that were detected at site 3D as separate events, based on changes in turbidity, chloride concentration, temperature, and dissolved oxygen concentration at site 3D, and daily records of reservoir level.

Broken lines give a schematised outline of each distribution.



Number of effective inflows
6
5
4
3
2
1
J
F
M
A
M
J
J
A
S
O
N
D

interflows and underflows overlap in the period April - June, and apparently in October - November also. The Inflow Register does not include any inflows having a maximum turbidity effect in October, so it is difficult to determine whether October is part of an overlap period. The Inflow register indicates that an underflow was initiated in mid-October 1976, which supports this assertion. The apparently equal representation of interflows and underflows in November is deceptive because, of the two recorded underflows, one had its maximum effect very early in the month (1/11/76; see Table 3.3), and the other resulted from the exceptional floods of 1961. Nevertheless, the smoothed distribution curves plotted in Fig. 3.9 are a reasonable approximation of the underflow and interflow periods that have resulted from the seasonal cycles of river and lake water temperatures. The minimum underflow period (June - August) predicted from Fig. 3.8 is based on the mean inflow temperatures, and includes 60% of the recorded underflows, and excludes 88% of the interflows (an average predictive success of 74%). If the underflow period is predicted from the minimum inflow temperatures (ie. May - September), then 85% of the recorded underflows are included and 76% of the interflows excluded, an average predictive success of 81%. Allowing for the possible contribution of turbidity, the 7 month underflow period (April - October) includes 90% of the recorded underflows, but excludes only 65% of the interflows, giving an average predictive success of 78%.

The best estimate from this model is that which compares the minimum inflow temperatures with the mean surface and deep-water temperatures at site 3D, giving a 5 month underflow period, May - September, with an average predictive success of 81%). It is apparent that this simple model, based on temperature measurements from four sites (Fig. 3.8), can be used to account for a high percentage of the recorded inflows, regardless of the relatively infrequent measurements of inflow temperatures (often only one reading per month), and the difficulty of accurately characterising river temperatures, particularly during floods, from such records. However, all the predictions from Fig. 3.8 are symmetrical around July. Optimum predictive success is, however, obtained from a 5 month period symmetrical about August (June - October), which includes 85% of underflows and excludes 88% of interflows, having an average predictive success of 87%. Although there is some change in the distribution of inflows if the inflow initiation dates are used, rather than the date of maximum recorded turbidity (see Table 3.3.), the overall predictive success is similar, except that there is no difference between the May - September and June - October underflow periods, both of which yield 87% average predictive success.

Comparison with Registered Inflows:-

Despite the success of this model, comparison of Fig. 3.8 with the inflow register (Table 3.3 ; Fig. 3.9), indicates that factors other than simply the inflow temperatures and the overall thermal gradient at site 3D. are significant in differentiating between interflows and underflows. Interflows occur in June, when even the mean river temperatures indicate underflow conditions, and have not been recorded for September or October, when the conditions indicate a much greater liklehood of interflow (Fig. 3.8).

A reasonable suggestion is that interflows are strongly influenced by the detailed thermal structure in the lake, ie. the steepness of the thermal gradient, and its depth in the water column. Consider the extension of the interflow period into June, when the thermocline is usually sharply defined and occurs relatively deep in the water column. A sharply defined gradient of 1.0 - 1.5°C (eg. June in Fig. 3.8), may be better at supporting an interflow than a less well defined gradient of similar overall magnitude (eg. September in Fig. 3.8). Perhaps more important, however, is the fact that, for a deep thermocline, an inflow travels further into the lake before lifting-off to become an interflow. Entrainment of epilimnetic water into a cold inflow must tend to decrease its density, favouring an interflow. In September and

October, when thermal stratification is being re-established, the thermal gradient tends to be less definite and much shallower. It is, therefore, conceivable that a cold inflow might plunge through the less sharply defined density barrier, especially as the inflow would encounter the thermocline closer to the inlet, where the inflow may have more velocity and would have entrained less of the warmer epilimnetic water.

Another possibility is that interflows occurring in September and October tend not to be transmitted to site 3D and are, therefore, not detectable in the present analysis. Several factors could contribute to this scavenging of the interflow by the actively mixed layer. First, the mixed depth is notoriously variable during the early stages of thermocline formation (cf. Darbyshire and Edwards 1972; Mortimer 1974). Second, it is possible that in this early warming phase, when there is often no clearly defined epilimnion, the metalimnion is a more actively mixing zone throughout its depth than is the case in subsequent months when an approximately isothermal epilimnion marks the region of most intense vertical circulation. Finally, the relatively shallow thermocline may ensure that an interflow is more accessible to wind-induced mixing than an interflow in a deeper thermocline with a similar temperature range. More detailed investigation of inflows in relation to the thermal structure, both vertical and longitudinal in Lake Burragorang, would help to answer some of the questions that arise from the previous speculative discussion which rested on a couple of general principles:-

1. That interflows travel in an active zone, when compared to the quiescent hypolimnion traversed by underflows. Periodic downward incursion of the epilimnion, combined with the effects of seiches, will tend to dissipate an interflow as it progresses into a lake. It is likely that these effects will vary with the seasonal cycle of thermal stratification in the lake. Further, whereas underflows are effectively channeled downslope to the outlet of a lake such as Lake Burragorang,

an interflow is subject to coriolis deflection and may not have the gravitational assistance imparted to an underflow following the old river course.

2. That, in the context of a warm monomictic lake, entrainment of water resident in the lake will tend to result in the inflowing water mass underflowing the layer from which the entrained water was derived. For example, an inflow colder than any water in the lake which entrains epilimnetic water may warm sufficiently to interflow, whereas entrainment of hypolimnetic water cannot contribute to an interflow, though it would reduce the density difference between itself and the inflow. Depending on the extent of the entrainment, a relatively warm inflow might entrain enough epilimnetic water for the turbidity induced density to cause it to interflow.

Seasonal patterns of inflow have recently been investigated for two warm monomictic lakes in New Zealand (Pickrill and Irwin 1982; Irwin and Pickrill 1982). A simple model, similar to that presented here, was reported by Pickrill and Irwin (1982) for lake Wakatipu. They found an underflow period from mid-April to early October (6 - 7 months) in a detailed investigation lasting twelve months. Interestingly, they found that for the remainder of the year a mixture of underflows and interflows occurred, with overflows during the summer. The continued ocurrence of underflows, throughout the year, was also found for Lake Tekapo (Irwin and Pickrill 1982) and contrasts with the behavior in Lake Burragorang where underflows have not been recorded during the summer (December - February). Lakes Wakatipu and Tekapo apparently have generally cold inflows which overflow and interflow in spring and summer as a result of large diurnal temperature ranges (Pickrill and Irwin 1982; Irwin and Pickrill 1982). This also produces cyclical interchange between interflows, underflows and overflows within a single day in summer. However, at least for Lake Wakatipu summer flood flows interrupt this diurnal pattern resulting in underflow (Pickrill and Irwin 1982). It should be noted that the present study of Lake Burragorang is confined to these larger scale inflow effects both in terms of time and inflow volume, so that the fine structure of daily variations is not visible.

A seasonal cycle of inflow has also been reported for Lake Mead (Hoffman and Jonez 1973), in which spring overflow was followed by summer interflow deepening through autumn to a winter underflow. Seasonal variation in the salinity of the Colarado River inflow played a significant role in this cycle however, in contrast to the cycle for Lake Burragorang. Upstream of Lake Mead, the more recently formed Lake Powell is reported to show a similar cycle (Johnson and Merritt 1979) and may well have altered the pattern in Lake Mead.

Weiss et al (1978) report that the Alpenrhine enters Lake Constance as an interflow from February to November and underflows only in December and January.

Clearly there are a wide variety of possible seasonal inflow cycles which depend on the specific character of the inflowing stream and its catchment, and the interaction of these factors with the cycles of thermal and chemical stratification in the lake.

CHAPTER 4

OXYGEN
STRATIFICATION

INTRODUCTION

Wetzel (1983) describes oxygen as "... the most fundamental parameter in lakes, aside from water itself". He goes on to say "... the solubility and especially the dynamics of oxygen distribution in lakes are basic to the understanding of the distribution, behaviour, and growth of aquatic organisms". Attendant upon the thermal stratification in lakes there develops a vertical zonation of oxygen concentration which reflects the restriction of wind-induced water circulation to the upper part of the water column and the relative isolation of the meta and hypolimnia from atmospheric oxygen. In these zones the demand for oxygen, from aerobic respiration and chemical oxidation, usually exceeds the rate of supply and they tend to become depleted of dissolved oxygen. Depending on the period of thermal stratification and the rate of oxygen depletion the hypolimnia of many lakes become anoxic. This has a variety of important consequences for the lake's biota, and in terms of lake manangement for water supply has prompted the development of direct manipulatory measures to prevent it.

In the classic account of natural lakes the cycle of oxygen stratification can be explained in terms of vertical processes (mixing and sedimentation of organic and inorganic material). However, in the present context the advective, effectively horizontal effects of inflows must also be considered. In this chapter a similar approach to that used for thermal stratification is adopted. The account moves from a general description of the annual cycle of oxygen stratification, based on monthly mean profiles (1961 - 1980), to a more detailed consideration of year to year variation and the role of advective processes.

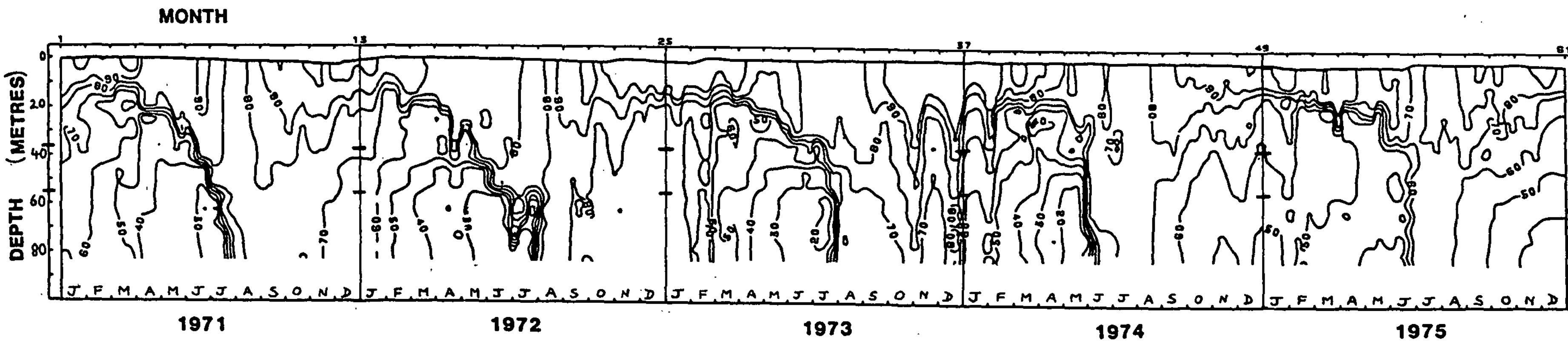
THE ANNUAL CYCLE OF OXYGEN STRATIFICATION IN LAKE BURRAGORANG

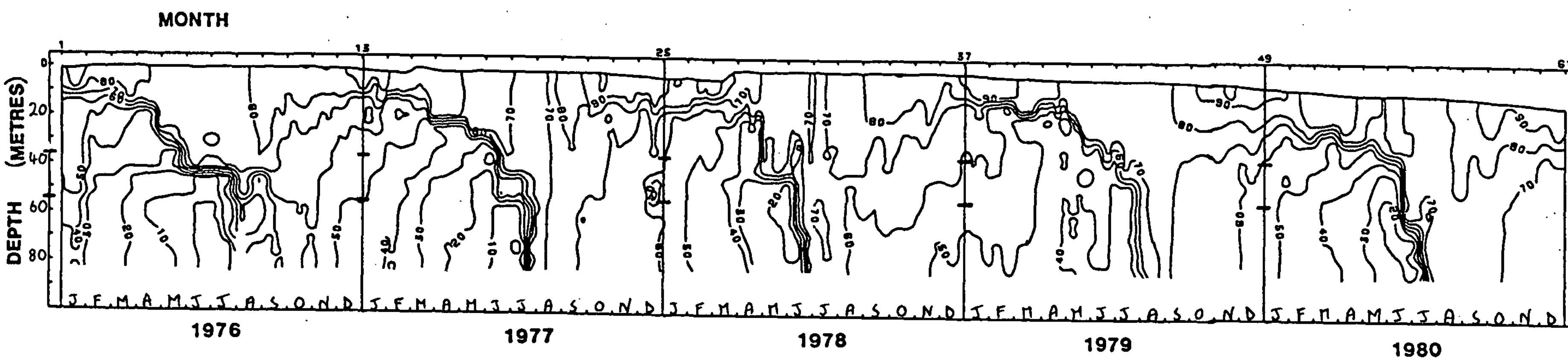
Some General Features

Isopleths of Oxygen percentage saturation for site 3D in the period 1961 - 1980 are given in Fig. 4.1. Several features of the oxygen stratification cycle are readily apparent from this diagram. Hypolimnetic anoxia is rarely observed in Lake Burragorang at its deepest point. Oxygen is commonly depleted to less than 30% of saturation but anoxia has been recorded only once, in 1962, following the unusually large and turbid inflows of November 1961. Table 4.1 contains a list of the lowest recorded oxygen concentration for each of the 20 years. Minimum annual records range from 0 to 35% of saturation (mean = 13%; SD = 10.6%; n = 20). Lake Eildon (Victoria; Powling 1980) and Lake Argyle (Western Australia; Imberger and Patterson 1979) are deep lakes that apparently, like Lake Burragorang, do not develop anoxic hypolimnia within the stratification period. In contrast, Lake Gordon (Tasmania: Steane and Tyler 1982) and Dartmouth Reservoir (Victoria: Welsh 1984) both exhibit anoxia in their deeper layers for some part of the year. In fact, Dartmouth has been described by Welsh (1984) as meromictic and is yet to display a convective overturn; hypolimnetic re-oxygenation has occurred through cold winter underflow. Both these reservoirs are more recent than Lake Burragorang, and they will probably tend towards lower hypolimnetic oxygen deficits with time. Data presented by Scribner (undated) for Burrinjuck Reservoir (New South Wales) indicates that anoxia is likely to develop; Scribner recorded a dissolved oxygen concentration of 0.1 mg l^{-1} on the 6/2/75, about two thirds of the way through the stratification period. Lake Barrington (Tasmnia; Tyler and Buckney 1974) has a deep meromictic pool, but the overlying hypolimnetic water also becomes anoxic during the stratification period. Shallower reservoirs, such as Lake Hume (Victoria; Croome 1980) and North Pine Dam (Queensland; King and Everson 1980), also

FIGURE 4.1

Dissolved oxygen (% saturation) isopleth diagram, 1961 - 1980. Isopleths are drawn at the following intervals:- 2, 5, 10, 20, 30, 40, 50, 60, 70, 80, 90, 100% saturation. Vertical lines are drawn at the end of each calendar year. Fluctuation of the lake surface is shown at the top of the diagram, indicating that the isopleths are drawn relative to a fixed datum (elevation), rather than relative to the lake surface. This more clearly shows the oxygen stratification in relation to the sub-surface offtake, the bounds of which are marked on the depth axes and on the dividing lines between calendar years.


MONTH
DEPTH (METRES)
J F M A M J J A S O N D
1971
1972
1973
1974
1975


MONTH
DEPTH (METRES)
J F M A M J J A S O N D
1976
1977
1978
1979
1980

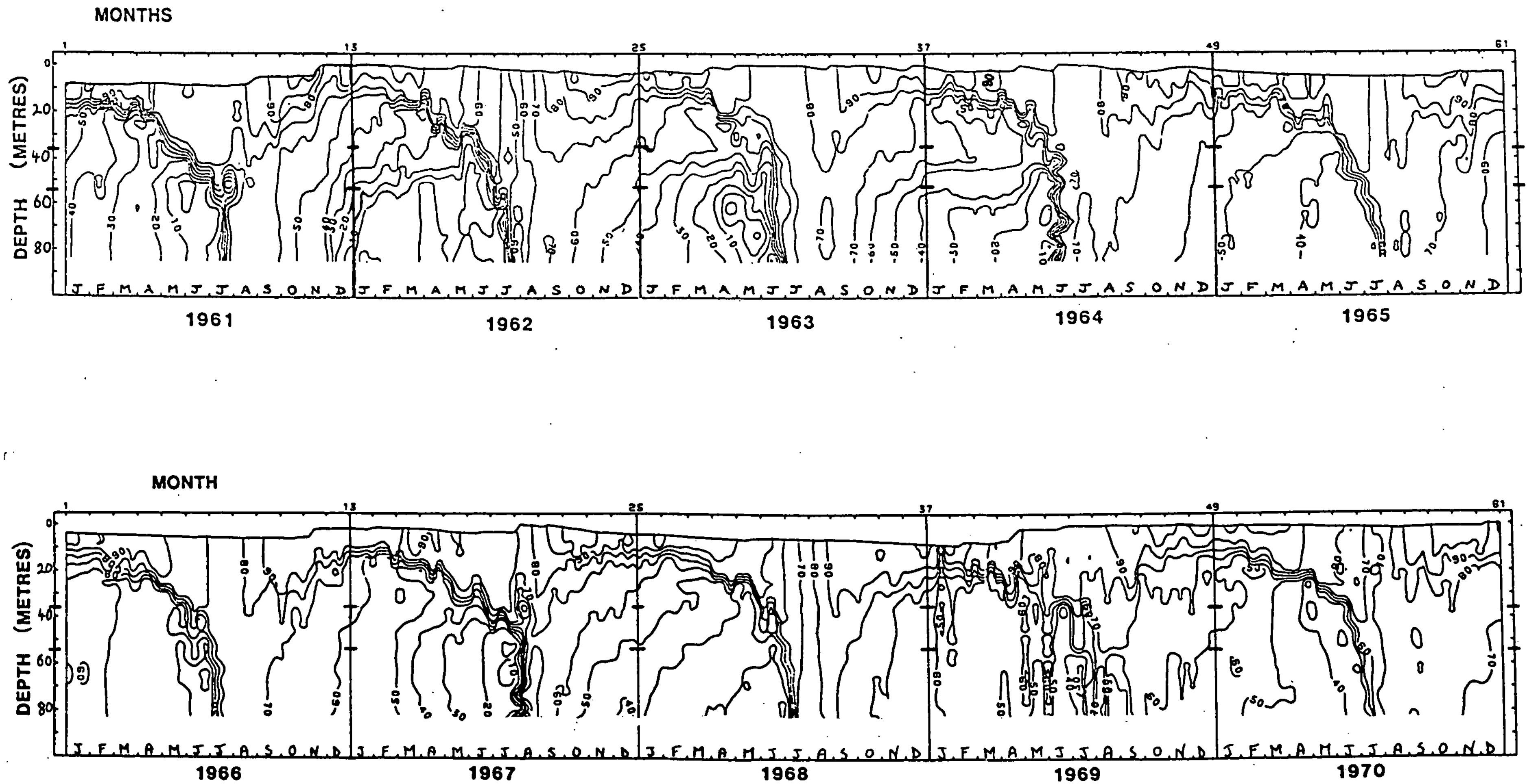

MONTHS
DEPTH (METRES)
1961
1962
1963
1964
1965
MONTH
DEPTH (METRES)
1966
1967
1968
1969
1970

Table 4.1 Annual minimum record of dissolved oxygen ($mg\ l^{-1}$ and % saturation), and sample depths, from 1961 - 1980.

Year	Absolute D.O. Minimum $mg\ l^{-1}$	% Saturation	Sample depth metres	Sample date(s)
1961	0.3	2.8	42.7 - 54.9	16 May : 30 May
1962	0.0	0.0	85.3	17 July
1963	0.1	1.0	61.0 : 73.2	23 April : 28 May
1964	0.2	1.9	61.0	2 June
1965	2.3	21.3 [a]	85.3 : 88.4	19 May : 21 July
1966	2.9	27.1	61.0	5 July
1967	0.5	4.7	61.0	1 August
1968	1.3	11.9	85.3	9 April
1969	2.4	22.9	61.0	15 July
1970	2.6	24.9	48.8 : 61.0	22 June : 29 June
1971	1.2	11.3	61.0	8 July
1972	1.7	16.3	61.0	24 July
1973	1.4	13.3	73.2	16 July
1974	1.2	11.6	61.0	6 May
1975	3.8	35.0	48.0	9 June
1976	0.3	2.8	72.0	21 June
1977	0.4	3.8	48.0 - 72.0	11 July
1978	0.7	6.7	60.0 : 72.0	6 July : 12 July
1979	3.0	30.0	12.0	23 April
1980	1.2	11.3	60.0	30 June

[a] 2.3 $mg\ l^{-1}$ equivalent to 21.1% Saturation at 85.3 m on the 19th of May.

develop anoxic hypolimnia each year.

The surface of Lake Burragorang is generally > 90% saturated with oxygen between September and March, a period of 7 months during which thermal stratification is initiated and reaches its peak prior to the autumnal deepening of the epilimnion. In this latter phase the progressive incorporation of oxygen depleted water, combined with the cooling of the epilimnion and therefore increased oxygen solubility, contribute to the decline of oxygen percentage saturation in the epilimnion. In winter, when vertical circulation is maximal, the water column is usually about 70% - 80% saturated with oxygen.

Peak values of surface oxygen percentage saturation occur between October and December (approximately) when the vertical extent of the mixed zone, and therefore the volume of ciculating water, tends to be least. This period is commonly associated with slight supersaturation of oxygen at the surface which probably results from both the warming of the surface layers, which decreases the solubility of dissolved gases, and algal production. However, the biomass (as chlorophyll-a) rarely exceeds 5 - 10 mg m^3 and the years of greatest algal populations do not coincide with those of greatest oxygen supersaturation. In 1973 supersaturation exceeded 150%, and such figures may be in error. Scribner (undated) provides data for a series of deep impoundments in New South Wales which indicates that supersaturation of up to 107% occurs at the surface of both Blowering and Burrinjuck Reservoirs at times when chlorophyll concentrations are less then 5 mg m^3. Hutchinson (1957) reported the findings of several authors who surveyed American, European, and Japanese lakes in the 1930's (Juday and Birge 1932; Ohle 1934; Lönnerblad 1931; Yoshimura 1938) and found from 7% - 45% of the lakes exhibited supersaturation. Yoshimura (1938) found that oligotrophic lakes ranged up to 120% of saturation, while mesotrophic and eutrophic lakes reached 126% and 166% of saturation respectively. Lake Burragorang may be considered mesotrophic at site 3D, based on the criteria for chlorophyll and total phosphorus concentration listed by Cullen and Rosich (1979). In most years supersaturation remains below c. 120% which is consistent with Yoshimura's (1938) findings.

The extent to which the circulating water column tends to equilibrate with the atmosphere can be estimated from the vertical isopleths in Fig. 4.1. These are comparatively infrequent, indicating that during the relatively brief periods of maximum vertical mixing (1 - 2 months) increases of more than about 10% of saturation are rare. In 1968 the increase may have been as much as 20% of saturation, bringing the water column to more than 80% of saturation by the end of the mixing period. Evidently the period of maximum circulation in Lake Burragorang is not sufficient to achieve saturation in the full volume of the lake.

Long Term Monthly Means of Dissolved Oxygen

The average cycle of oxygen stratification can be determined from the long term monthly means of oxygen concentration at various depths, calculated for the study period (see Methods, Chapter 2). Fig. 4.2 (Oxygen percentage saturation) and Fig. 4.3 (Oxygen concentration, mg l^{-1}) show the progression of these values with time. As a result of the long term averaging the diagrams may be regarded as circular, with January succeeding December. Profiles of the dissolved oxygen percentage saturation data are given in Fig. 4.4, and as mg l^{-1} (with standard deviations) in Fig. 4.5. Month to month difference profiles for oxygen percentage saturation appear in Fig. 4.6. Representative profiles (1 month^{-1}; from individual sample days at site 3D) for eight years of the study period are contained in Appendix 2.

Epilimnion:-

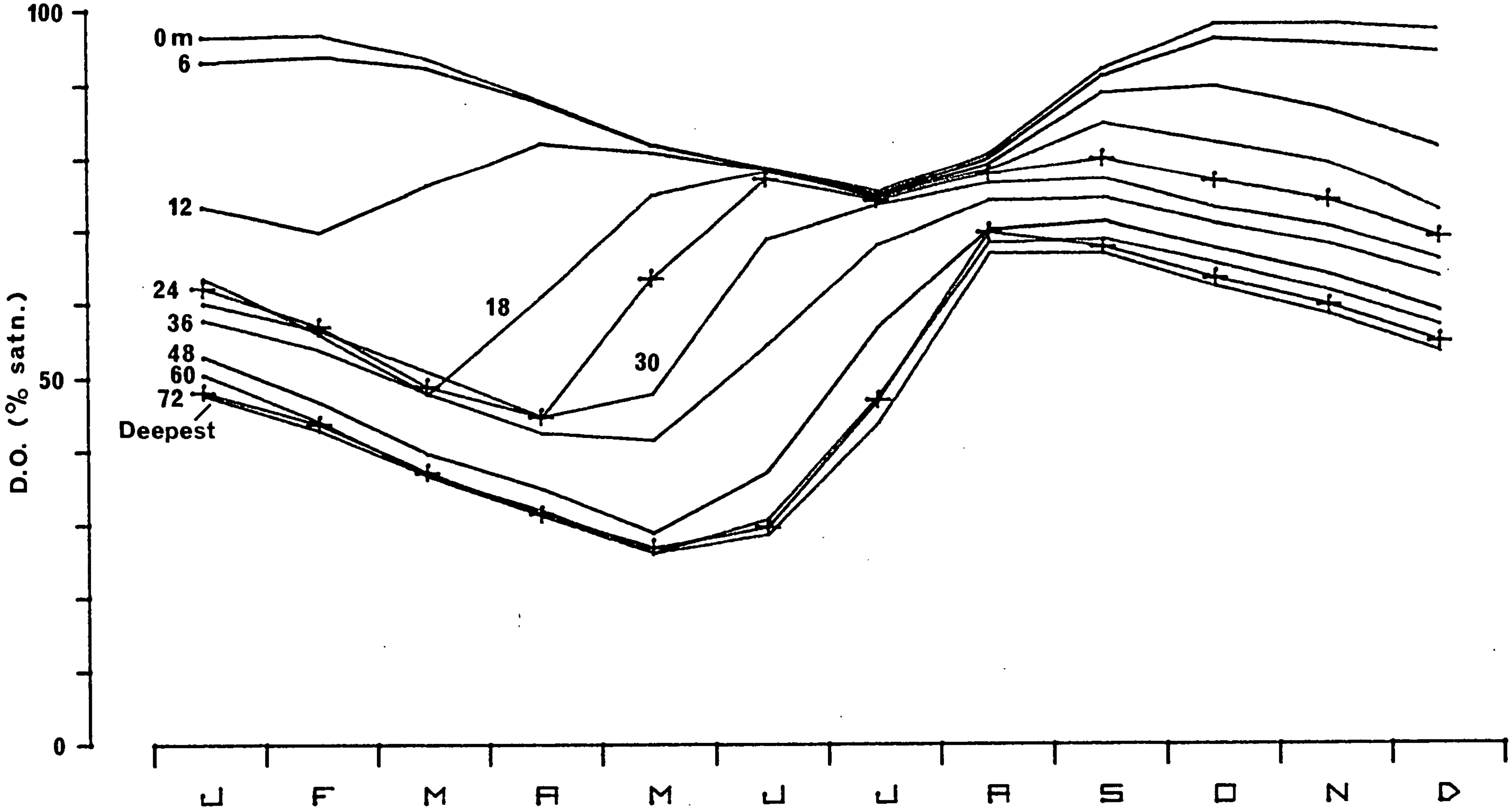
The behaviour of the epilimnion, which naturally varies in vertical extent with season, can be characterised in terms of oxygen concentration by the

FIGURE 4.2

Seasonal progression of twenty year monthly mean dissolved oxygen (% saturation) for 11 standard depths (see Materials and Methods, Chapter 2). The same data *are* presented as profiles in the Fig. 4.4. The formation of a metalimnetic oxygen trough, during the first 3 - 4 months of the year, is evident from the crossing of some lines (particularly 18 m).

Symbols:-

24 m and 72 m (+)



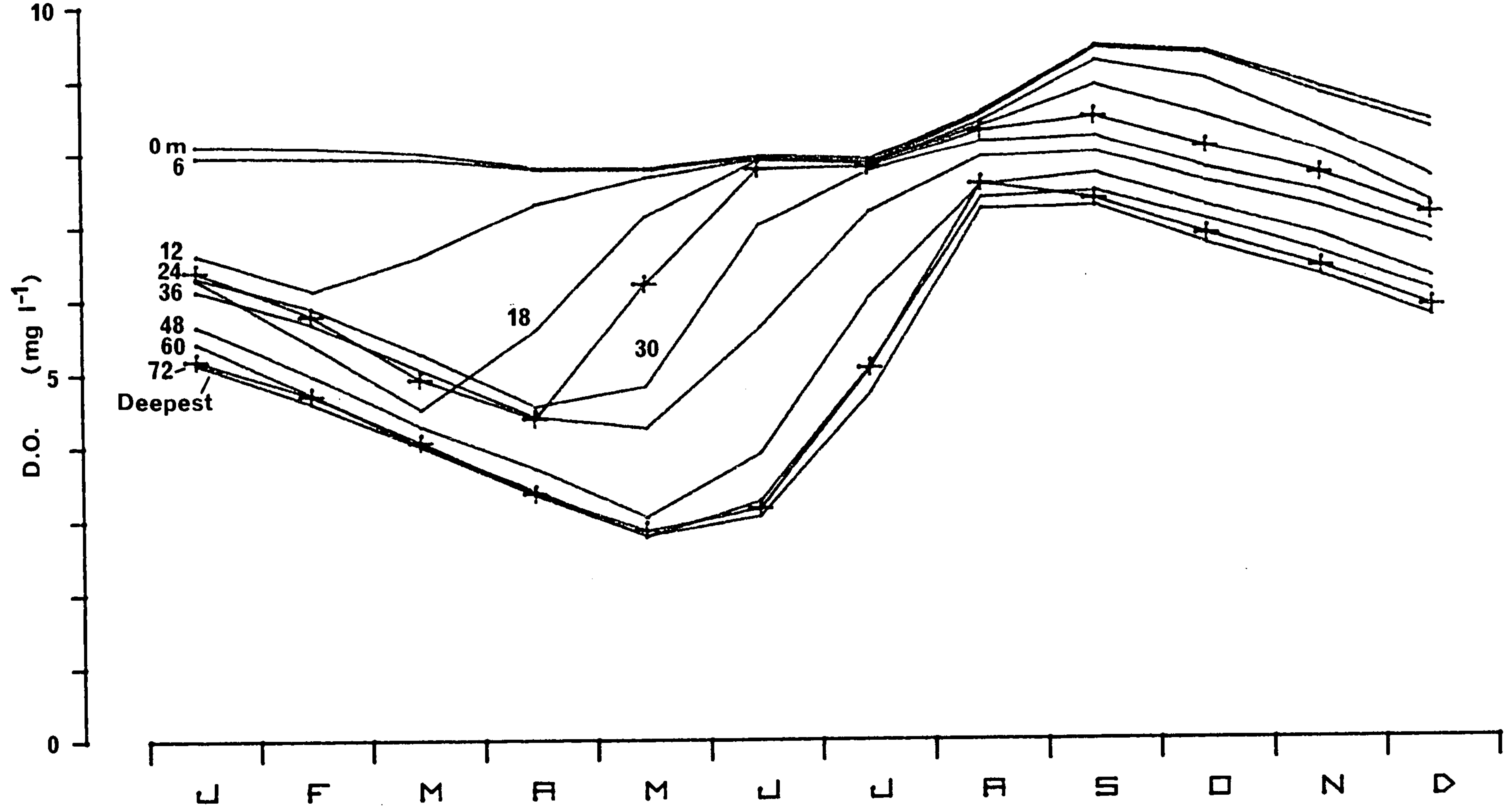
100
50
0
D.O. (% satn.)
0 m
6
12
18
24
30
36
48
60
72
Deepest
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 4.3

Seasonal progression of twenty year monthly mean dissolved oxygen (mg l^{-1}) for 11 standard depths (see Materials and Methods, Chapter 2). The same data *are* presented as profiles in the Fig. 4.5, with standard deviations. The formation of a metalimnetic oxygen trough, during the first 3 - 4 months of the year, is evident from the crossing of some lines (particularly 18 m).

Symbols:-

24 m and 72 m (+)



10
5
0
D.O. (mg l-1)
0 m
6
12
24
36
48
60
72
Deepest
18
30
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 4.4

Profiles of monthly mean dissolved oxygen (% saturation). Each profile is derived from the full twenty year dat set, according to the procedure outlined in the caption to Fig. 4.5 (see also Materials and Methods Chapter 2).

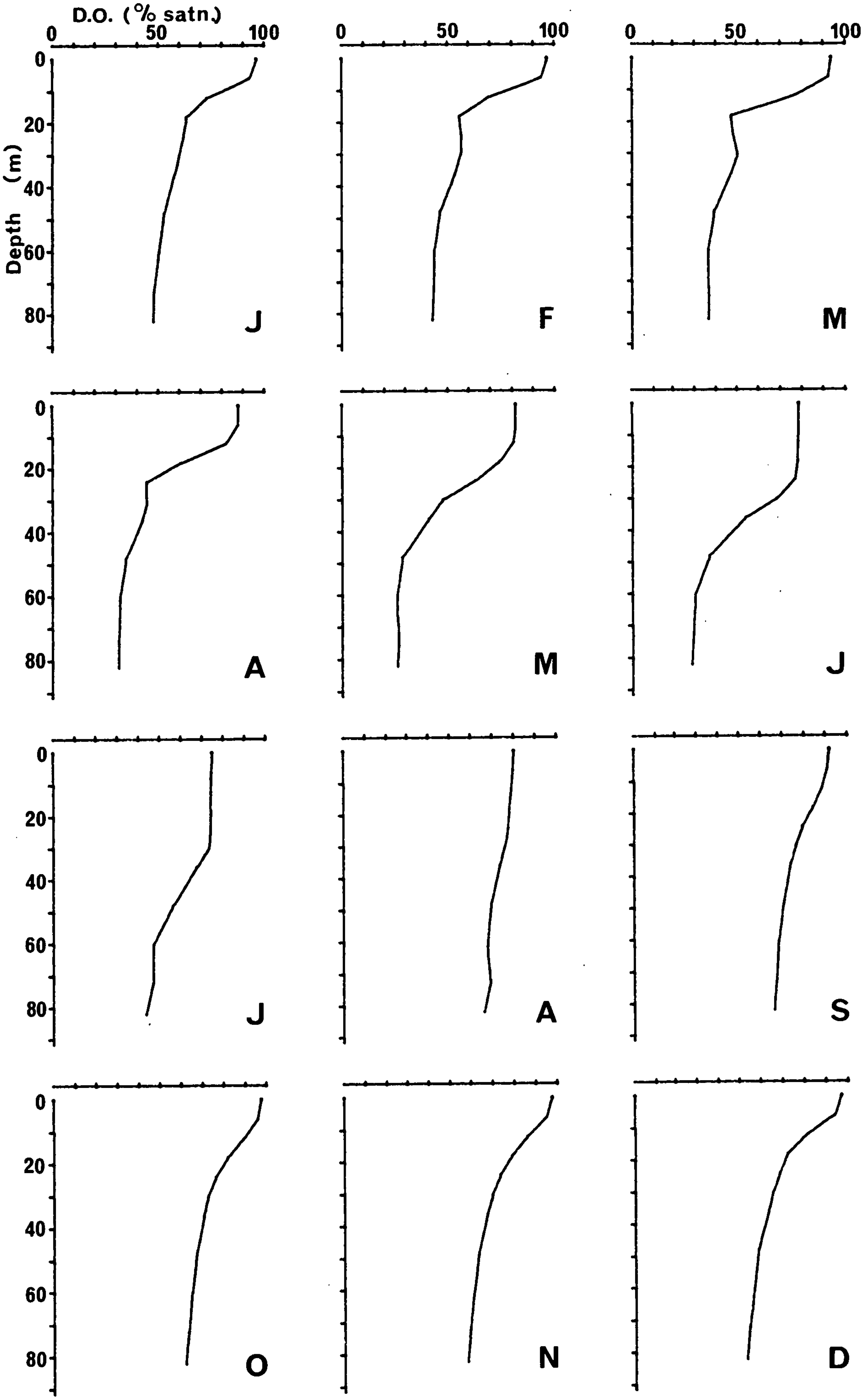

D.O. (% satn.)
0
50
100
Depth (m)
0
20
40
60
80
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 4.5

Profiles of monthly mean dissolved oxygen (mg l^{-1}; 1961 - 1980), calculated in two stages:- (see Methods, Chapter 2)

1. The monthly mean, for a particular depth below the water surface (eleven depths in all), is calculated for each of the twenty years, yielding 12 (months) x 20 (years) x 11 (depths) means, each with an n of from 1 - 5.
2. Then, the mean of the 20 individual monthly means is calculated, yielding 12 profiles, each with 11 depths.

The first of these steps, is designed to remove the effect of changing sampling frequency, so that the weekly sampling of the early years did not overshadow the 2 weekly or monthly samples of later years. The standard deviation for any point (ie January, at 6 m) describes the dispersion of the 20 individual monthly means (one for each year 1961 - 1980), about the overall mean.

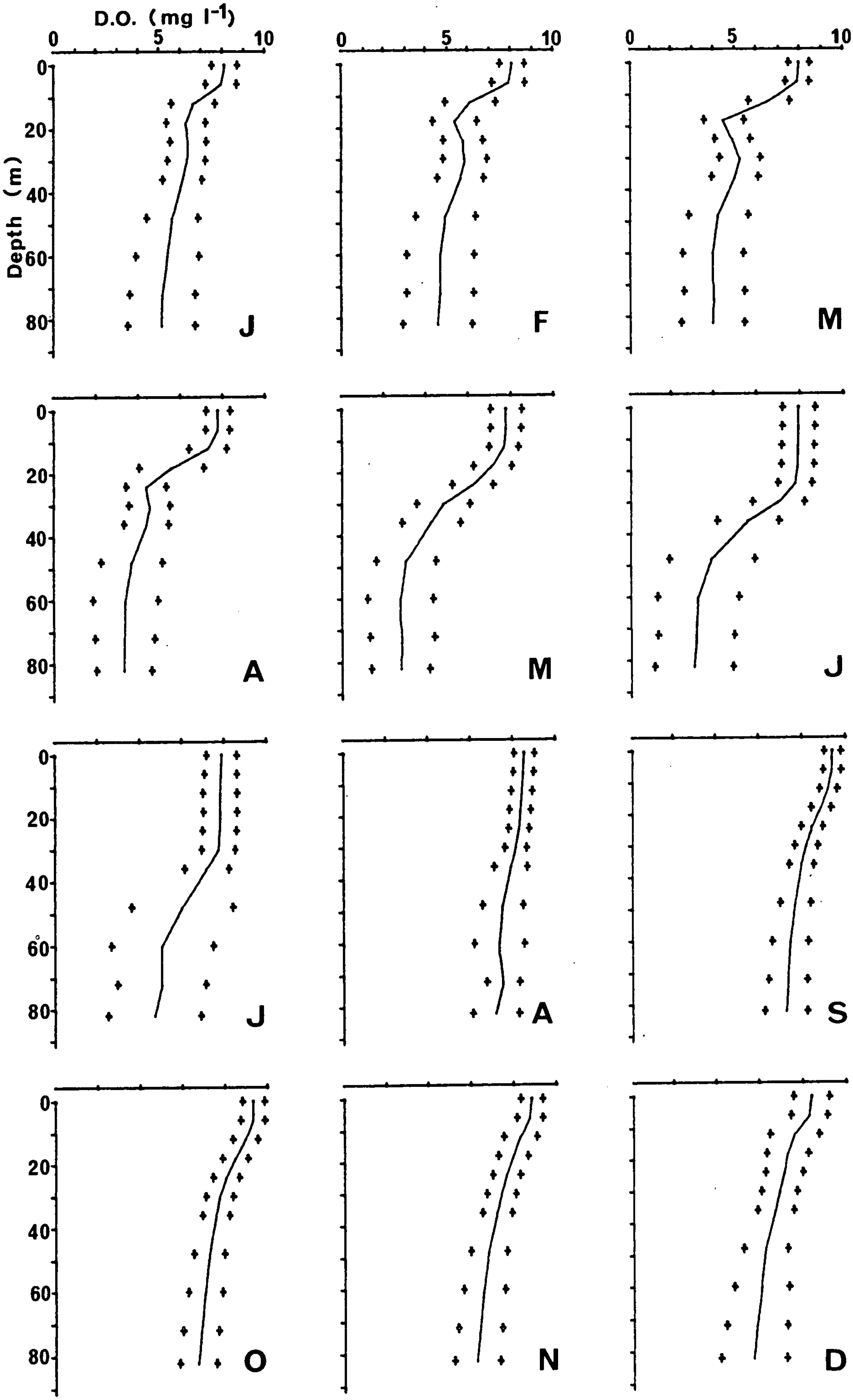
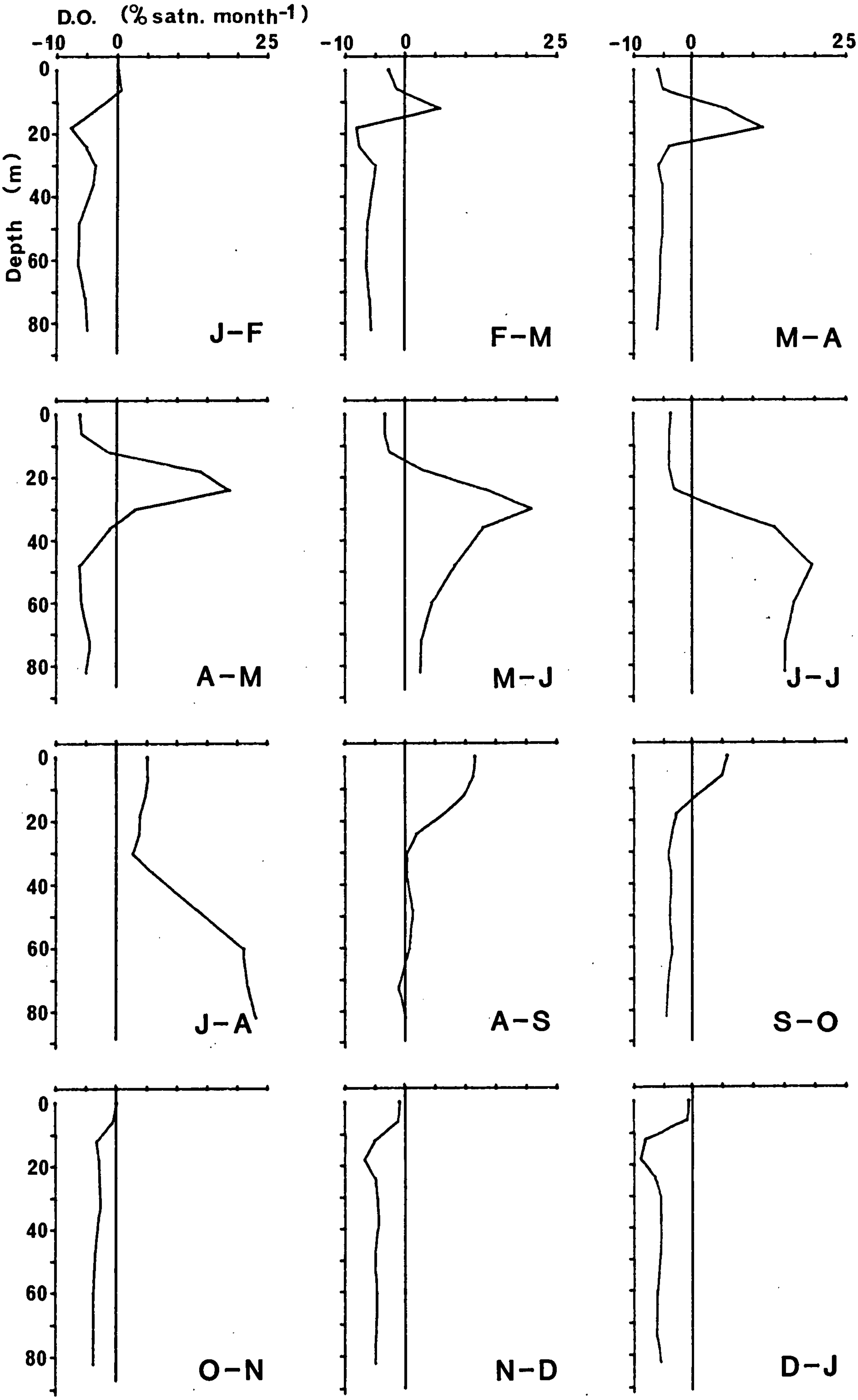

D.O. (mg l-1)
0
5
10
Depth (m)
0
20
40
60
80
J
F
M
A
M
J
J
A
S
O
N
D

FIGURE 4.6

Dissolved oxygen (% saturation) difference plots, showing the change in dissolved oxygen between successive months, for each of the sampled depths. A vertical line on each plot marks zero change from one month to the next. These differences were accumulated for each individual year, and a mean taken, so that each point of the profile is a mean of n = c. 20, and can be assigned a standard deviation, though these are not plotted here. This representation highlights the changes that occur between the monthly profiles of dissolved oxygen, giving a clearer picture of the dynamic aspects of the oxygen stratification.



D.O. (% satn. month-1)
-10
0
25
Depth (m)
0
20
40
60
80
J–F
F–M
M–A
A–M
M–J
J–J
J–A
A–S
S–O
O–N
N–D
D–J

region from 0 m to 6 m. These two sample depths remain within c. 5% of saturation of one another for the whole annual cycle. In January and February this zone is c. 95% saturated, declining to c. 74% of saturation in July when the epilimnion extends to 30 m at least (Figs. 4.2 and 4.4). This decline is not quite linear, having its maximum between April and May. In the same period (February - July) Fig. 4.3 shows that the oxygen concentration (mg l^{-1}) has only a slight trough in April and May (minimum annual concentration, in May, is 7.8 mg l^{-1}) indicating that the decline in oxygen percentage saturation is primarily a result of epilimnetic cooling and is only slightly affected by the progressive incorporation of oxygen depleted water from the deeper layers, or that the rate at which oxygen enters the water column just balances this effect. Epilimnetic oxygen percentage saturation increases from July to the annual maximum in October/November (c. 97% of saturation). This increase results from both increasing water temperature, and therefore decreased oxygen solubility, and from increased dissolved oxygen concentration (mg l^{-1}; Fig. 4.3). This latter effect is initially associated with the period of maximum vertical circulation, but between August and September the reformation of the thermocline and attendant reduction in mixed depth confines further increases in dissolved oxygen (mg l^{-1}) to the epilimnion only (c. 0 - 20 m; Figs 4.3 and 4.5). Consequently, oxygen stratification is actively initiated in the epilimnion, rather than simply by oxygen consumption in the meta- or hypo-limnia. From October to January epilimnetic (0 - 6 m) oxygen concentration (mg l^{-1}) declines, while oxygen percentage saturation peaks in October/November and remains high until February (Fig. 4.2). The decline in oxygen concentration (mg l^{-1}) occurs as the surface mixed layer warms and equilibrates to the lower oxygen solubility. As previously mentioned, in certain years this period is characterised by oxygen supersaturation in the surface layers.

Metalimnion:-

Profiles of oxygen percentage saturation (Fig. 4.4) and the diagrams of oxygen (percentage saturation, and mg l^{-1}) versus time (Fig. 4.2 and Fig. 4.3) show a negative heterograde oxygen profile in February and March, and perhaps April, after which the metalimnetic trough is destroyed by the deepening mixed zone. The month to month difference plots (Fig. 4.6) clearly indicate the tendency for a negative heterograde profile to develop from October - November through to February - March when it has its maximum development (Figs 4.4 and 4.5). This form of oxygen profile is relatively common in stratified lakes (Hutchinson 1957), and is reported for large impoundments such as Lake Powell (Utah - Arizona; Johnson and Merritt 1979) and Lake Mead (Nevada - Arizona; Hoffman and Jonez 1973). According to Wetzel (1983) it is less frequent than its opposite, the positive heterograde profile, but development of a positive heterograde profile usually occurs in response to algal oxygen production in the metalimnion and requires a ratio of euphotic depth to mixed depth of about 1 or greater. In view of the high turbidity frequently associated with Australian lakes (Kirk 1977a, 1977b; Ganf 1980), and the often low ratio of euphotic to mixed depths (Rosich 1983; see also Ferris 1977), it is reasonable to expect negative heterograde oxygen profiles to be most common in Australian lakes and reservoirs. Welsh (1984) reported a negative heterograde profile from Dartmouth Reservoir (Victoria). A variety of mechanisms have been suggested to explain the occurrence of the metalimnetic oxygen minimum. These include:- (1) The decomposition of material sedimenting from the epilimnion, proceeding at a greater rate in its early stages and in the comparative warmth of the metalimnion (Hutchinson 1957). (2) Oxygen depleted water is moved horizontally toward the centre of the lake from the sediment/water interface where the metalimnion contacts the lake sides, particularly when the metalimnion lies adjacent to a gently sloping region of the lake basin and is, therefore, exposed to a greater sediment area per unit volume than the hypolimnion (Wetzel 1983). (3) The advective effects of turbid, oxygen demanding, interflows (Ruttner 1963), and the entrainment of reduced chemical species (chemically oxygen demanding) from the hypolimnion by interflows (Nix 1981). Virtually all of these processes may operate in Lake Burragorang, with the probable exception of the entrainment of reduced chemical species as the hypolimnion is so rarely anoxic. The advective influence of turbid metalimnial inflows is anticipated to be of considerable importance however.

Fig. 4.6 also shows a downward moving wave of positive concentration change between February - March and May - June which is associated with the deepening mixed zone. The crest of this wave of re-oxygenation moves from 12 m (February - March) to 30 m (May - June), after which it becomes a part of the general re-oxygenation of the lower water column, caused by either underflow or overturn; the two are not differentiated in the averages derived from the full data set. A similar feature was noted for the thermal profiles, but whereas the zone of re-oxygenation remains obvious in the May - June difference plot, the heating wave has been obliterated by then (Chapter 3, Fig. 3.4). The crests of these two waves do not exactly coincide.

Hypolimnion:-

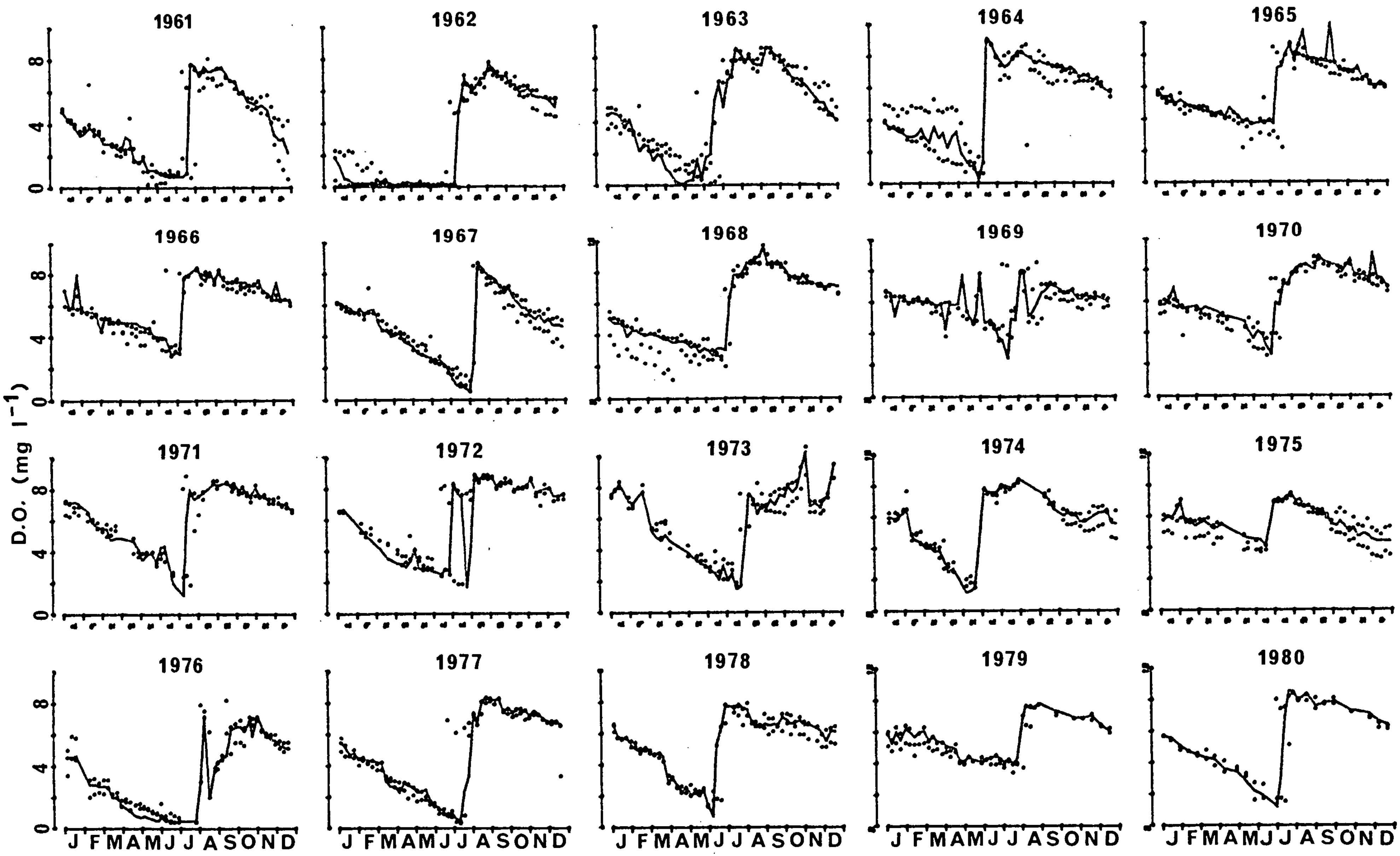
In the 20 yr averaged plots (Figs 4.2, 4.3, 4.4, and 4.5) the hypolimnetic oxygen concentration can be characterised from samples taken below about 48 m, though the hypolimnion commonly has a greater vertical extent than this, extending to within 24 m of the surface for c. 5 months of the year (Figs 4.4, 4.5, and Chapter 3, Fig. 3.3). Oxygen depletion definitely commences by October and continues until May, after which the effect of underflow or vertical mixing is to increase the hypolimnetic oxygen percentage saturation to its maximum, in August and September (c. 70% of saturation, Fig. 4.2; and 7.5 mg l^{-1}, Fig. 4.3). The 20 yr monthly means (Fig. 4.2 and Fig. 4.3) give the impression of a gradual re-oxygenation lasting from June to August/September. This results from the averaging of the relatively rapid periods of re-oxygenation that typify individual years, but occur at different times. The progression of oxygen concentration (mg l^{-1}) for individual years, in the region below 48 m, is shown in Fig. 4.7. The figure indicates that much of the hypolimnetic oxygenation in any one year occupies a period less than a month, and also demonstrates the variation in timing of this event.

Fig. 4.2 and Fig. 4.3 show that, overall, volumetric hypolimnetic oxygen depletion is approximately linear outside of the four month period when oxygen replenishment can affect the deeper layers of Lake Burragorang. The rate of oxygen depletion is approximately 5.1% month^{-1} (0.58 mg l^{-1} month^{-1}) for the 8 months from September to May. This can be divided into two sections of slightly different rate, 4.2% month^{-1} (0.51 mg l^{-1} month^{-1}) between September and December, increasing to about 5.6% month^{-1} (0.63 mg l^{-1} month^{-1}) from January to May. Division of the depletion period into two is an arbitrary procedure and the tendency for the rate of oxygen depletion to increase is almost certainly a smooth, rather than step-like, progression. Welsh (1984) reports similarly increasing rates of oxygen depletion over periods up to 6 months in Dartmouth Reservoir (Victoria). For Lake Burragorang the increasing rate of hypolimnetic de-oxygenation may result from the gradual sedimentation of spring and summer algal crops and/or the slight increase in hypolimnetic temperature during the stratified period, but inflow probably also plays a part by bringing in oxygen demanding material from the catchment.

The long term average rates of volumetric oxygen depletion permit some comparison between Lake Burragorang and other lakes. Vollenweider and Janus (1984) report volumetric hypolimnetic oxygen depletion rates (VHDR) for 21 European and North American lakes (mainly derived from the OECD Cooperative Programme on Eutrophication). These data, corrected to a uniform temperature of 4°C, have a mean of 0.787 mg l^{-1} month^{-1} (Standard deviation = 0.651; range = 0.09 - 2.11) and a median of 0.54 mg l^{-1} month^{-1}. Lake

FIGURE 4.7

The annual progression of dissolved oxygen concentration (mg l^{-1}) at site 3D for individual years (1961 - 1980). Samples from 60 m below the surface are joined by a line, while the dots represent samples taken between 48 m and the deepest sample (nominally c. 82 m), except for 1961 in which the dots represent samples from 42 m to the deepest sample.


1961
1962
1963
1964
1965
1966
1967
1968
1969
1970
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980
D.O. (mg l^{-1})
0
4
8
J F M A M J J A S O N D

Burragorang's long term mean (0.58 mg l^{-1} month^{-1}), corrected to 4°C (cf Vollenweider and Janus 1984), is 0.32 mg l^{-1} month^{-1}, less than either the mean or median of the lakes, including the North American Great Lakes and some of the deep European lakes (L. Maggiore, L. Léman, and Bodensee), in Vollenweider and Janus' (1984) study. This may be taken as evidence of the moderate to low productivity of Lake Burragorang, which is also indicated by the fact that oxygen is virtually never completely exhausted despite the long stratified period characteristic of the lake.

It is interesting to apply one of Vollenweider and Janus' (1984) models to Lake Burragorang. Their first equation predicts volumetric hypolimnetic oxygen depletion rates from chlorophyll concentration, euphotic depth, and mean depth. The mean of annual mean chlorophyll values for the period 1970 - 1980 is 2.2 mg m^{-3} at site 3D. For the full study period the mean of annual mean Secchi disc depths is 4.3 m (and for the period 1970 - 1980 incidentally). If it is assumed that the Secchi depth can be used estimate the euphotic depth by applying a constant multiplier of from 1.5 (Garman 1983) - 2.0 (Carlson 1977), then Vollenweider and Janus' (1984) model predicts a range of mean values from 0.347 - 0.448 mg l^{-1} month^{-1}, although incorporating the Standard Error (0.131) increases the range (0.216 - 0.579 mg l^{-1} month^{-1}; all at 4°C). Their regression tends to slightly over-estimate the measured VHDR for site 3D (0.32 mg l^{-1} month^{-1}; corrected to 4°C), although it does fall within the range of their prediction ± 1 Standard Error. Overall, the measured VHDR (long term mean) seems to lie at the lower end of the rates predicted by the model, but within the bounds of the models range. This does not, however, indicate that the model is necessarily suited to the prediction of year to year variation in VHDR for Lake Burragorang, a warning sounded by Vollenweider and Janus (1984) themselves.

Profile Overview:-

This section is in part a summary of the preceeding individual sections on the epi-, meta-, and hypolimnia and also provides an overall account of the seasonal oxygen cycle from the profiles of the 20 year monthly means and differences (oxygen percentage saturation; Fig. 4.4 and Fig. 4.6), which separate depth and time facets and more clearly show the processes that affect the three limnetic regions during the course of the year.

In August during the annual period of maximum vertical circulation the oxygen profile (Figs 4.4 and 4.5), though not quite iso-oxic, is generally monotonic and is the only profile where the standard deviation envelope overlaps from top to bottom (ie enabling a vertical line to be drawn that stays within 1 standard deviation of the mean; Fig. 4.5). This is almost true of the July profile where the hypolimnetic means have the greatest variability of any for the year, but a discontinuity remains between 30 m and 60 m. This accords with the long term thermal profile (July; Chapter 3, Fig. 3.3), but shows the discontinuity more clearly.

From August to December the oxygen profiles have a similar form to those of the corresponding thermal profiles (Chapter 3, Fig. 3.3). The difference plots (August - September and September - October, Fig. 4.6) show the beginning of hypolimnetic oxygen depletion and the increase in oxygen in the upper 20 m of the water column. They also show the accelerating trend in hypolimnetic depletion rate with time. The Difference plots (November - December to February - March, Fig. 4.6) show the continued hypolimnetic depletion and the tendency towards more rapid depletion in the metalimnion (c. 10 - 20 m), which in turn results in the negative heterograde profiles of February, March and possibly April. Deepening of the mixed zone is evident from March to July, and a zone of positive oxygen change has its peak at 10 m in February - March moving down to 30 m in May - June (Fig. 4.6). In the twenty year averages this zone coincides with the metalimnion, but in individual years probably represents the leading edge of the mixed layer. This zone may still exist in the June - July difference plot, with a peak at 48 m, but by this time the entire hypolimnion is being re-oxygenated, a process that continues in the July - August difference plot (Fig. 4.6). Consequently there is no month to month difference plot that shows a uniform oxygen increase at all depths, as might be expected from a freely circulating water column. This does not provide evidence for the absence of overturn, but does indicate that such periods are usually very short (c. 1 month). This estimate is shorter than that concluded from the 20 yr thermal difference plots where the July - August plot was almost uniformly negative with depth (Chapter 3, Fig. 3.4), and illustrates the value of oxygen profiles in determining particularly the onset of overturn.

During the summer period of heating and the early phase of mixed layer deepening the zone of negative oxygen change (in the metalimnion) overlaps with a zone of positive heat change; this is particularly obvious in the February - March difference plot (Fig. 4.6, and Chapter 3, Fig. 3.4). As heat is being transferred downward, and presumably also oxygen, the rate of oxygen depletion must exceed that of its supply. This emphasizes the difference between oxygen, which is actively consumed at all depths in the lake, and heat which is subject to relatively insignificant loss except at the surface. Although no distinction is drawn here between convective and advective processes, this phenomenon is later shown to be mainly associated with the combined effect of inflow and outflow (advective processes).

VARIATION OF THE ANNUAL CYCLE OF OXYGEN STRATIFICATION

A closer examination of the oxygen percentage saturation isopleths (Fig. 4.1) provides some greater detail of the year to year variation in Lake Burragorang. It is also instructive to compare the oxygen isopleths with the temperature isopleths (Chapter 3, Fig. 3.1).

A general feature is the tendency for the oxycline to become sharpened, and therefore most distinct (tightly bunched isopleths), in the autumnal cooling phase as the mixed zone deepens. The oxycline usually begins its descent gradually, accelerating towards the final re-oxygenation of the hypolimnion which appears as a vertical group of isopleths in almost every year (Fig. 4.1). The timing of this event varies over a period of 4 months (May - August). Re-oxygenation of the hypolimnion occurs annually (despite the lack of complete overturn in many years), most frequently in July, and in response to either underflows or overturn. The variation in timing is related to inflow in that re-oxygenation prior to July is invariably brought about by a cold, oxygenated underflow while re-oxygenation by overturn is confined to July and August, in which months the re-oxygenation can result from either process. Examples of wet years with early re-oxygenation are 1963, 1964, 1974, 1975 and 1978 when re-oxygenation occurred between May and early July (Fig. 4.1). The effect of inflows, in this respect, is to increase the variability of the system. Related to this last point is the depth of the mixed zone prior to the rapid re-oxygenation of the hypolimnion. This is evident from the depth of the oxycline and varies between years over a depth range of some 40 m (c. 30 m - 70 m; below FSL; Fig. 4.1). Part of this variation is undoubtedly a result of the lower water levels typical of dry years, and this could account for up to c. 10 m since the isopleths are drawn relative to a fixed datum. However, much of it results from the ability of underflows to short-circuit the convective overturn while the epilimnion is still relatively shallow.

Metalimnetic troughs of dissolved oxygen are evident in the first 3 - 4 months of many years, sometimes to the extent of a closed isoline as in February 1970 and March/April 1973 when closed circles of 50% saturation occur at about 20 - 25 m below FSL. These troughs are most clearly seen from the individual oxygen and temperature profiles given in Appendix 2. Localised troughs of oxygen percentage saturation are not confined to the metalimnion.

Two other types of localised depletion occur at site 3D in certain years.

The first is produced by inflows, when some of the old hypolimnetic water is displaced upwards by an underflow, leaving an oxygen profile with a trough between two oxygenated zones. A particularly clear example of this is August 1967 (Fig. 4.1) where a trough is centered at c. 36 m below FSL, overlying the near vertical isopleths that mark the hypolimnetic re-oxygenation caused by a sudden inflow (cf Inflow Register, Chapter 3, Table 3.3). Another example comes from early August 1961 where a bulge of isopleths remains, centered at c. 50 m below FSL, after an inflow beginning in July. Although this upward displacement of old hypolimnetic water occurs in other years (ie 1973, 1974, 1975), it is less obvious from the isopleth diagram (Fig. 4.1). The same phenomenon was noted by Steane and Tyler (1982) in Lake Gordon (Tasmania), a lake in which snow-melt contributes to cold, winter underflows. Similarly, Welsh (1984) reports the same sequence of events during an underflow into Dartmouth Reservoir (Victoria).

The second type of localised oxygen depletion is found in the hypolimnion during Autumn and early winter, when the minimum oxygen concentration in the water column occurs at mid-depth rather than at the deepest sampled depth. Obvious examples of this include May - June 1961, April - May 1963, May 1964, July 1967, June 1973, and May - June 1980 (Fig. 4.1). This behaviour probably results from the operation of the offtake structures at Warragamba Dam, particularly the fixed level hydroelectric (HEPS) offtake which has the greatest capacity, and will be treated in more detail later.

The relationship between thermal and oxygen stratification

A comparison of the oxygen and temperature stratification patterns (Fig. 4.1, and Chapter 3, Fig. 3.1) shows that there is good agreement of the fine structure, such as the vertical movements of the thermocline and oxycline, in most years. For example, in those years which show marked vertical fluctuations of the thermocline between October and December (1965, 1975, 1977, and 1978), this movement is also evident in the oxygen percentage saturation isopleths. There are two exceptions to this, in 1969 and the latter part of 1973. This lack of co-ordination between temperature and oxygen data probably indicates problems with water sampling or analysis at the time of data collection (see Materials and Methods, Chapter 2).

Another comparative feature of these two diagrams is a clear tendency for the oxycline to occupy the upper portion of the metalimnion, beginning just below the actively mixed epilimnion. This is most readily seen in the first few months of the year (ie 1961, 1964, 1967, 1979; Fig. 4.1). Although this may be regarded as inherent in the tendency toward the development of metalimnetic oxygen troughs, it points to the value of oxygen measurements as indicators of the depth of the actively mixed zone especially in summer and autumn, and accords with Hutchinson's (1957) observation that the lower metalimnion is in some respects more closely related to the hypolimnion than it is to the upper metalimnion. A related observation is that the oxycline frequently rises during the heating period, from a depth of 20 - 30 m at the time of its earliest formation, to within about 10 m of the surface in the middle of summer. This is apparent from the 70% and 80% saturation isopleths (Fig. 4.1) and gives the overall impression of a series of adjacent arches. This is taken as evidence of the progression of the mixed zone from greater to lesser vertical penetration as the surface waters are heated in summer, despite the fact that a rising thermocline is not readily apparent from the temperature data.

Both of these last general observations can be inferred from the long term average plots (Fig. 4.3, and Chapter 3 Fig. 3.2). In these figures the sample interval (6 m) is constant for the upper 36 m of the water column, and the vertical spacing between the plotted lines of oxygen concentration (mg l^{-1}; Fig. 4.3) and temperature (°C; Fig. 3.2) therefore indicates the gradient of the profile in this region, for each month. Fig. 4.3 shows the upward progression of the oxycline from 18 - 24 m in September to 6 - 12 m in November, where it remains until February. In contrast, the thermocline lies between 12 and 18 m from September to March, with the exception of November when it occupies the 6 - 12 m band. Between September and November/December the gradients of both oxygen and temperature are diffuse so that the distinctions drawn here are fine ones. However, if more than one depth interval is considered, then the region of maximum oxygen gradient is from 12 - 24 m in September and about 6 - 24 m in October/November, but by December it is established between 6 and 12 m indicating that the mixed zone has retreated upward from the depth of its formation.

Imberger and Hebbert (1980), in a detailed account of diurnal changes in the thermal stratification of Wellington Reservoir (Western Australia), comment that "Experience suggests that we would find, early next morning, a slightly warmer layer uniformly mixed to 8 metres. Note that the bulk thermocline is at a depth of about 12 metres, so for these sunshine dominated days, the bulk WML (well mixed layer) is itself tending to retreat.". Although their observations were more detailed, both in terms of time and also of depths sampled, they suggest that the well mixed layer and the main thermocline need not be exactly coincident. I believe that this may be of importance with respect to regions typified by oceanic thermal profiles where it is frequently difficult to find a distinct epilimnion, especially on the basis of profiles taken near the middle of the day and with fairly coarse depth resolution. In these circumstances it is usual to rely on the depth of the thermocline to estimate the mixed depth, since it is apparent that heat is still transferred to significant depths despite the apparent lack of an epilimnion. To illustrate this point, it is relatively easy to build up a typical summer profile for Lake Burragorang, in which the thermocline remains within one sample interval (6 m), from a sequence of profiles in which the depth to which heat is transferred is progressively shallower (Fig. 4.8). The data presented here is sufficient only to propose that the mixed depth in Lake Burragorang may retreat upwards in the period of rapid heating, and that dissolved oxygen data presents the best indicator of the mixed depth in a system where oceanic thermal profiles may have no conveniently isothermal layer from which to infer a well mixed region.

In January and February when the gradients are more strongly developed the region of maximum oxygen gradient clearly lies between 6 and 12 m, while the thermocline is found between 12 and 18 m, illustrating the generality of the observation that the oxycline tends to overlie the thermocline at the peak of stratification in Lake Burragorang.

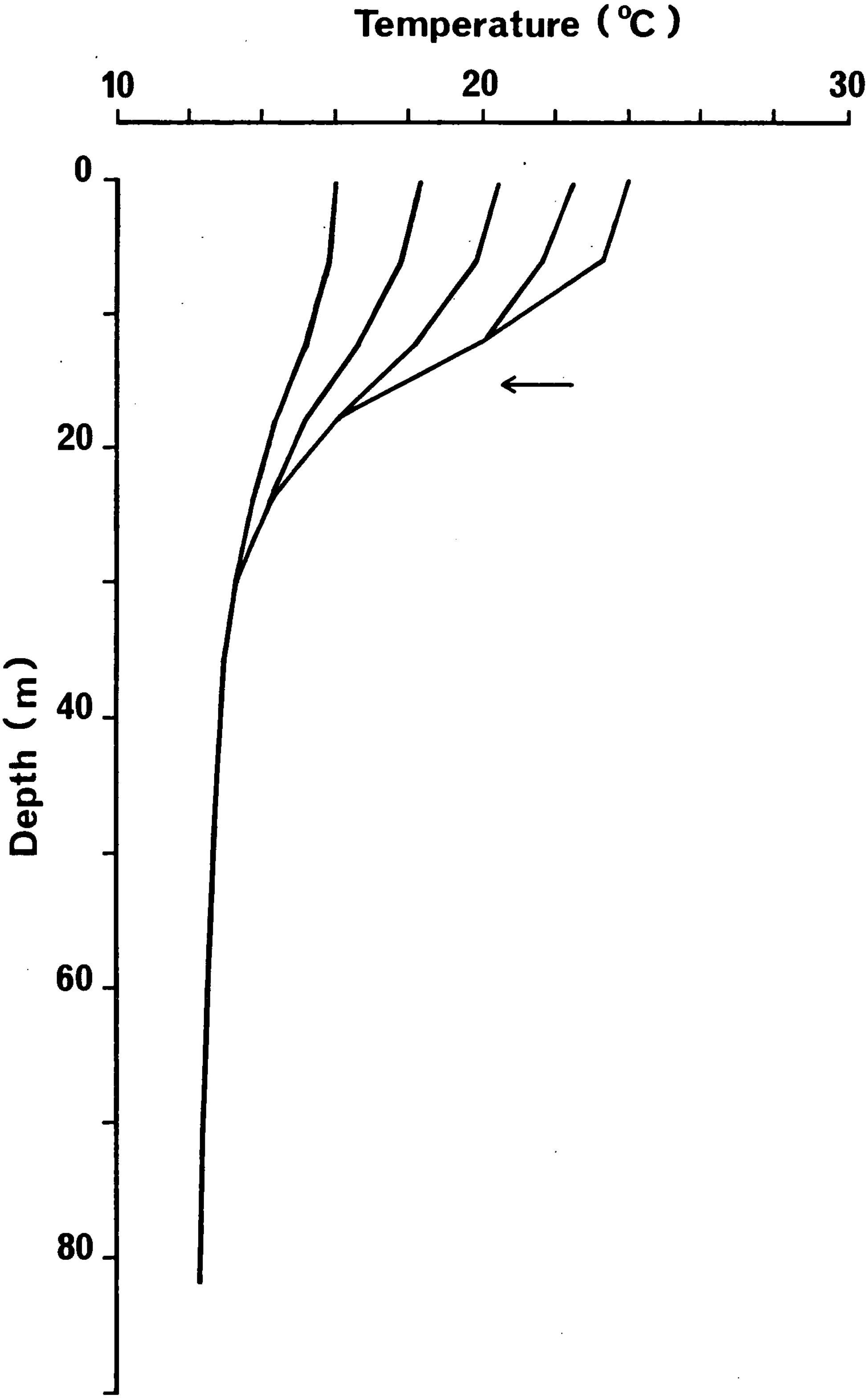
MID-DEPTH OXYGEN MINIMA IN THE HYPOLIMNION

The existence of localised oxygen minima in the hypolimnion was mentioned in the previous section with reference to Fig. 4.1. From Table 4.1 it is evident that the minimum annual record of hypolimnetic oxygen concentration is often found at mid-depth in the hypolimnion, rather than near the sediment/water interface. This is unexpected regardless of whether the dominant contribution to hypolimnetic oxygen demand comes from the sediment/water interface or from the water itself. The phenomenon is most obvious in the autumn and early winter, before the re-oxygenation by inflow or overturn, and examples from 1961, 1963, 1964, 1967, 1973, and 1980 have been mentioned previously (Fig. 4.1).

A characteristic of reservoir hydrodynamics which probably explains this behaviour is the occurrence of horizontal currents caused by subsurface withdrawal from the lake. Various studies in the laboratory and field (ie Johnson 1974; Imberger and Hebbert 1980) have provided a theoretical basis for the understanding and prediction of the vertical zone affected by

FIGURE 4.8

A schematic representation of the development of a typical summer thermal profile in Lake Burragorang, showing the possibility of maintaining the main thermocline (marked with an arrow) within a single 6 m depth interval while the depth to which heat is transferred becomes progressively shallower.



Temperature (°C)
10
20
30
0
20
40
60
80
Depth (m)

withdrawal structures in density stratified reservoirs. Thermal stratification restricts the layer affected by withdrawal through the dam wall, by suppressing vertical water movement, and enhancing horizontal movement (Wunderlich 1971; Wunderlich and Elder 1973). The width of this layer depends on the positioning of the offtake structure and the form of the thermal stratification. Generally, however, water is taken from a narrow band at about the same level as the offtake (Wunderlich 1971). Fiala (1966) points out that this selective withdrawal can lead to the interleaving of water layers, from different parts of the reservoir, adjacent to the dam. If de-oxygenation is further progressed at the upstream (shallower) sites in the reservoir, then a band of relatively less oxygenated water may be drawn into the water column near the dam producing the localised troughs found at site 3D (Fig. 4.1). Fiala (1966, 1979) reported this sequence of events in Klícava reservoir (Bohemia), and the tendency toward earlier development of oxygen depletion at shallower sites in reservoirs is relatively common. May (1978) gives evidence of this from Burrinjuck Dam (N.S.W.), and Scribner (undated) presents data for eight irrigation impoundments in N.S.W., five of which (including Burrinjuck) showed a greater de-oxygenation at the shallower upstream sites in late summer. This can also be established for Lake Burragorang, from the comparison of dissolved oxygen profiles at site 3D (adjacent to Warragamba Dam) and the Bend site (c. 14 km upstream of the dam). The progression of oxygen concentration (mg l^{-1}) at 60 m for these two sites, is shown for 1964, 1966, 1967, 1973, and 1976 in Fig. 4.9. By autumn to early winter the Bend tends to develop an oxygen concentration about 1 - 2 mg l^{-1} lower than is found for the same depth at site 3D. This is apparent for 1964, 1966 and 1967 (Fig. 4.9), and even in 1973 and 1976, the oxygen concentration is generally lowest at the Bend.

Warragamba Dam has two subsurface outlets which are in regular use, the water supply offtake operates at any depth between 0 and about 60 m while

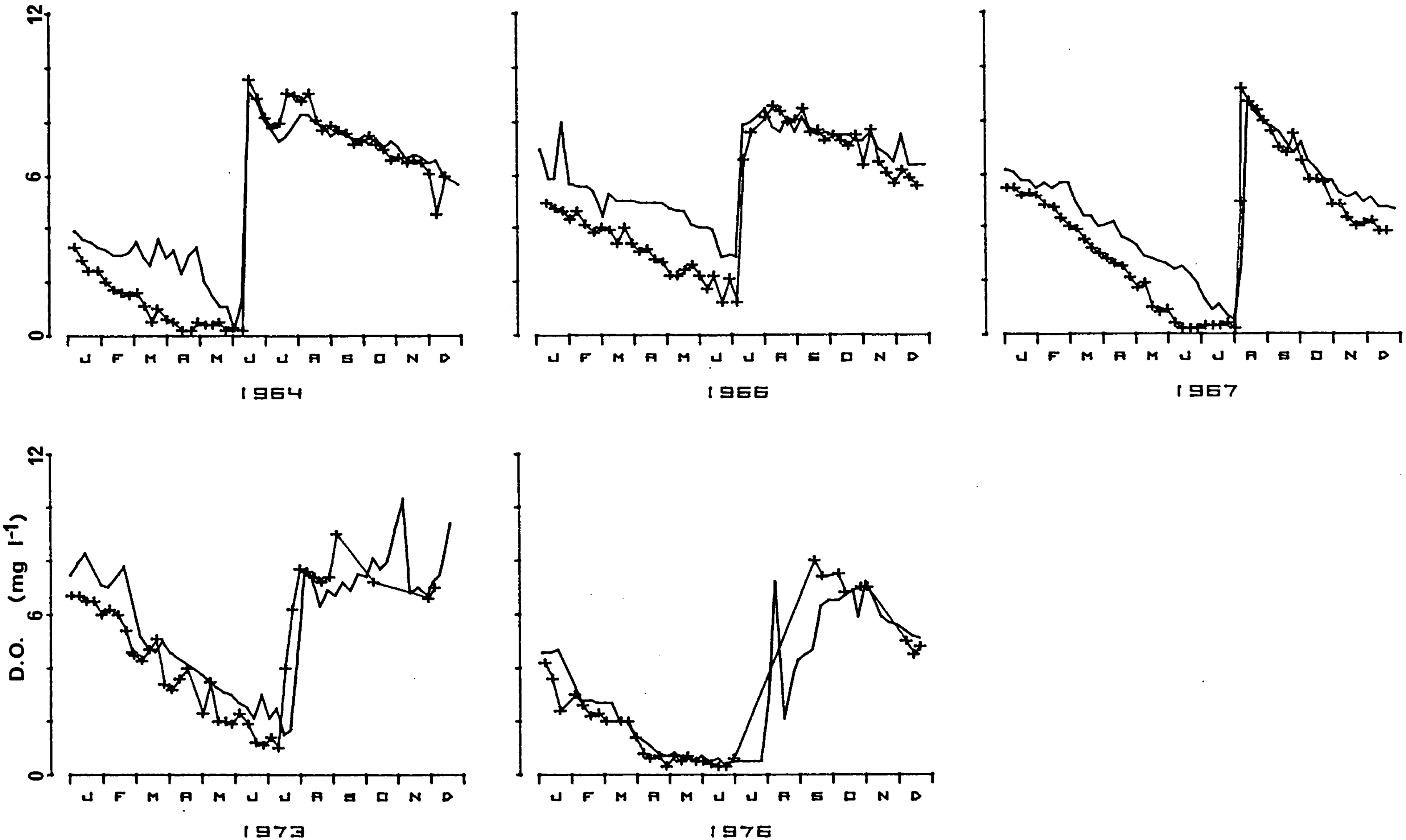
FIGURE 4.9

A comparison of dissolved oxygen concentration (mg l^{-1}) taken from 60 m (below the surface) at sites 3D, and Bend, plotted against time. The general tendency for the upstream site (Bend, c. 14 km upstream of 3D) to develop lower oxygen concentrations at a given depth is most obvious in 1964, 1966 and 1967.

Symbols:-

3D (·)

Bend (+)



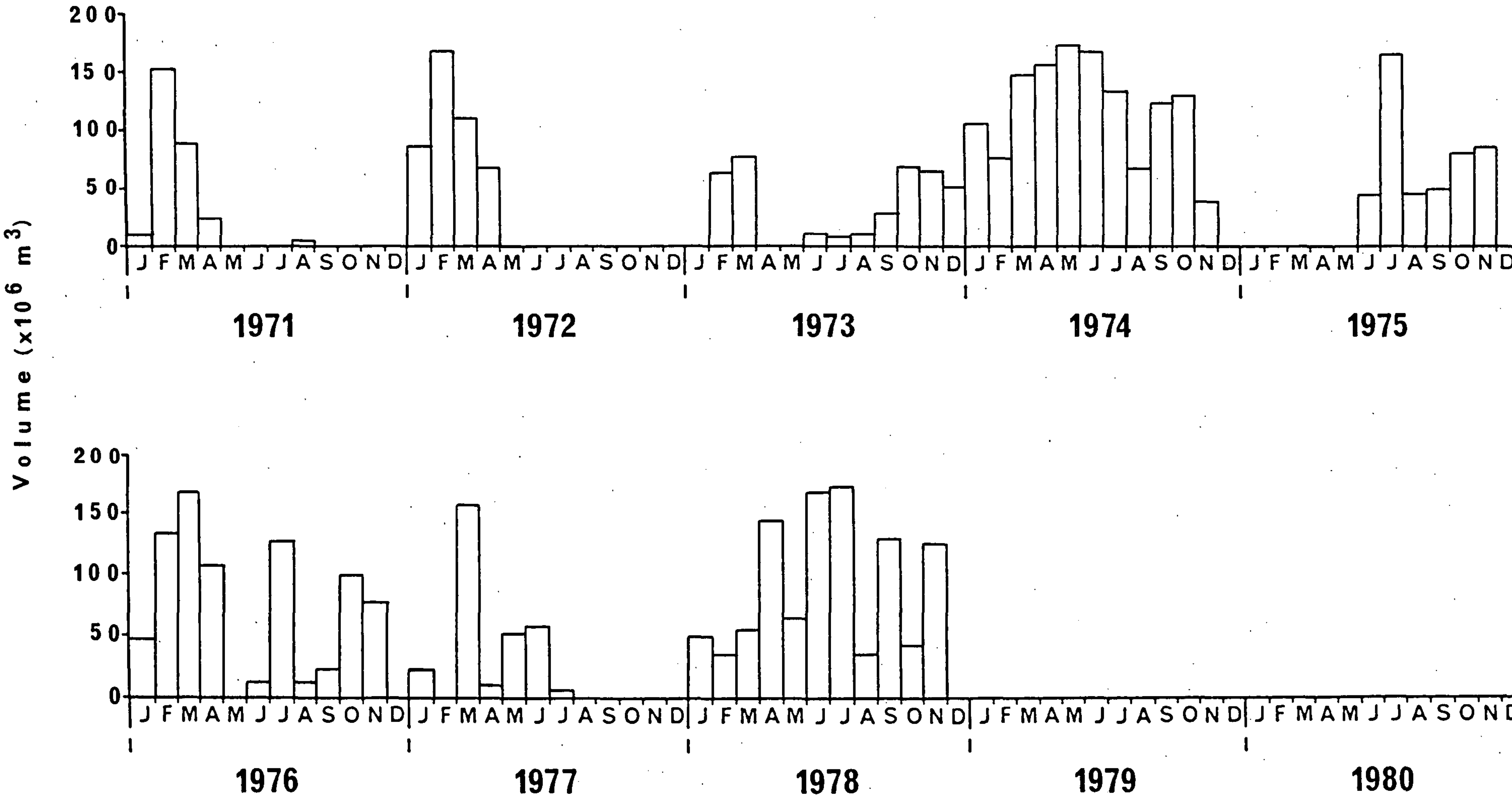
D.O. (mg l-1)
0
6
12
J F M A M J J A S O N D
1964
1966
1967
1973
1976

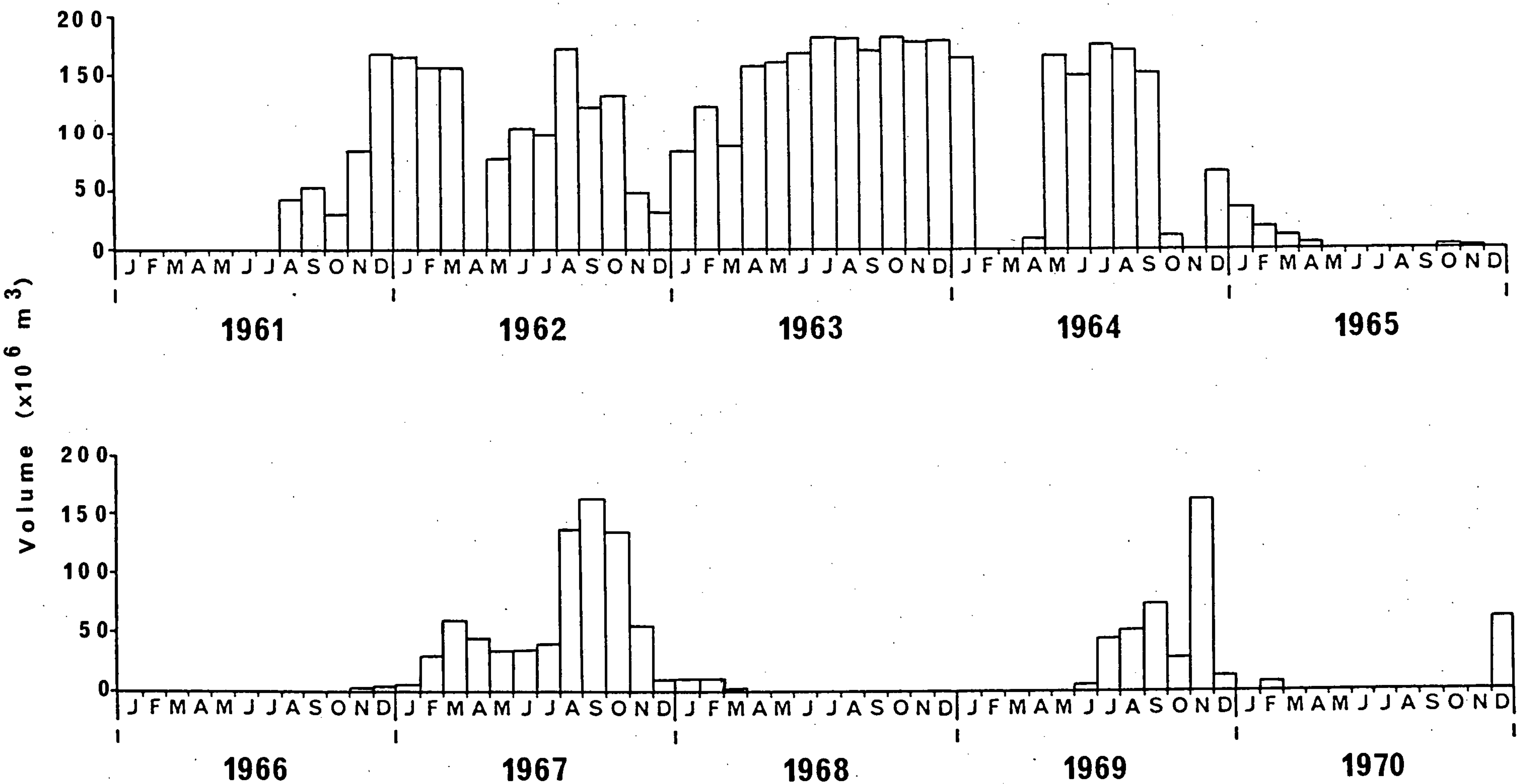
the Hydro Electric Power Station (HEPS) offtake, which has more than twice the capacity of the water supply offtake, is a fixed structure centred at 44.5 m (below FSL). The HEPS offtake stands on a sill protruding from the dam wall (3 - 4 m) and it is 18.3 m high (35.4 - 53.6 m below FSL). Within the period for which outflow records are available (August 1961 - December 1980), the maximum monthly withdrawal through these structures was 181 10^6 m^3 (HEPS offtake) and 61 10^6 m^3 (Water supply offtake). If transposed to annual figures these represent 106% and 36% of the total lake volume (at FSL) respectively. In the present context I will consider primarily the HEPS offtake, because of its greater capacity and the fact that it effects the hypolimnion for much of the year. Comparison of the oxygen percentage saturation isopleths (Fig. 4.1; the HEPS offtake level is marked) and the monthly HEPS outflow volumes (Fig. 4.10) shows a general concordance of the mid-depth hypolimnetic troughs with periods of high volume discharge through the HEPS offtake; for example in April - June 1963, and May - June 1964. Interestingly, the troughs usually formed near the bottom of the HEPS offtake or below it. Perhaps more importantly the figures show that in some years, when the HEPS was not operated significantly, the localised hypolimnetic depletion did not eventuate (ie 1965 and 1968). An obvious exception to this is 1980, when a mid-depth trough formed in May and June, and the slight trough formed in June 1966 also occurred when the HEPS was not operating. These will be further considered later.

In the context of a limited comparison between the Bend and 3D sample sites, 1964 gives the clearest indication of the apparent ability of the HEPS offtake to draw water from further upstream into the vicinity of the dam wall. Fig. 4.9 shows the annual progression of dissolved oxygen (mg l^{-1}) at 60 m (below the surface) for these two sites in 1964; there is a definite convergence of the two plots in May to early June prior to the June flood (see Inflow Register; Chapter 3, Table 3.3). This coincides with a sudden increase

FIGURE 4.10

Monthly total outflow volume (x 10^6 m^3) through the sub-surface HEPS offtake, which extends from 35.4 m to 53.6 m below FSL. The data covers the period from August 1961 - December 1980. The volume scale is 1/8th of that used for the total inflow (minus evaporation) diagram (Chapter 3, Fig. 3.6). I have no data for the first 7 months of 1961, but the HEPS was probably in use, though at a low volume. Operation of the HEPS offtake is closely tied to the occurrence of inflow by management policy. Consequently, no use was made of the HEPS in 1979 or 1980 as the water in storage was reserved for supply to Sydney, during these drought years. Used continuously, and at full capacity, the HEPS is capable of withdrawing the volume equivalent of the lake (at FSL) in about 1 year.


Volume (x10^6 m^3)
200
150
100
50
0
J F M A M J J A S O N D
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980


Volume (x10^6 m^3)
200
150
100
50
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

(to > 150 10^6 m^3 $month^{-1}$; Fig. 4.10) in the volume of water withdrawn through the HEPS offtake, and despite the variation in oxygen concentration at site 3D (March - April) seems to mark a definite change in the rate of oxygen decline at site 3D. Fig. 4.11 shows the sequence of dissolved oxygen (mg l^{-1}) profiles at the two sites in May and June. From profiles showing quite different hypolimnetic oxygen concentrations and lacking any mid-hypolimnetic trough (5 - 6th May), the profiles converge in the region adjacent to the HEPS offtake, and the 3D profile tends to develop the mid-depth trough (18 - 19th May). The profiles taken on the 25th and 26th of May show the plots touching within the depth range directly affected by the HEPS offtake. Profiles from the 1st - 2nd and 9th of June show the oxycline oscillating to within the region of the HEPS offtake and are not consistent in relation to the oxygen concentration at site 3D.

The evidence presented here cannot be considered unequivocal, but the there is undoubtedly reason to assert that the HEPS offtake could cause the observed mid-hypolimnetic trough. This is especially so if the capacity of this outlet is considered in relation to the volume of water stored at and below the level of the HEPS offtake. About 344 10^6 m^3 of water, or 17% of the total lake volume, lies below the top of the HEPS offtake (35.4 m below FSL). If the thermocline can be considered the upper boundary of the zone directly influenced by the HEPS offtake, then, as the thermocline descends in autumn the HEPS offtake draws from a smaller and smaller volume of the hypolimnion, tending to the limiting condition of withdrawal from about 17% of the lake's volume. In which case the HEPS offtake is capable of subtracting this entire volume of water in less than 2 months. This also suggests that the HEPS offtake may have some effect at less than its full withdrawal capacity, for example, the development of a mid-hypolimnetic trough in 1967 apparently coincides with HEPS withdrawal volumes around 40 - 50 10^6 m^3 $month^{-1}$. This, in turn, raises the question of the potential effect of the

FIGURE 4.1.1

Comparative profiles of dissolved oxygen concentration (mg l^{-1}) at sites 3D and Bend for 1964. Samples were usually taken on successive days.

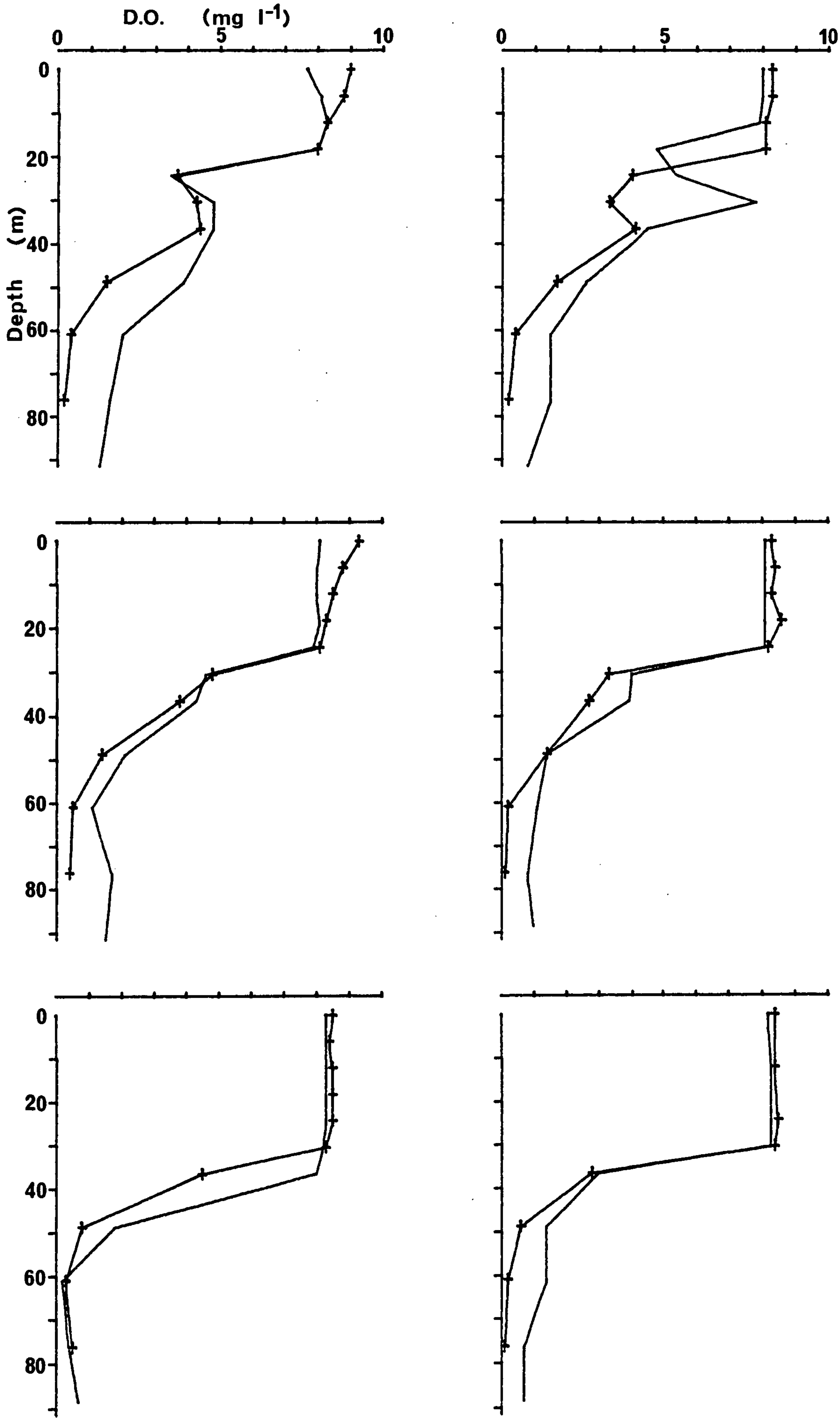
Symbols:-

3D (·)

Bend (+)

Sample Dates:-

3D	5/5/64	12/5/64
Bend	6/5/64	13/5/64
	19/5/64	26/5/64
	19/5/64	25/5/64
	2/6/64	9/6/64
	1/6/64	10/6/64



D.O. (mg l^{-1})
0
5
10
Depth (m)
0
20
40
60
80

water supply offtake. In 1980 the HEPS offtake was not operated, yet a mid-hypolimnetic trough developed in June; the temperature of the subtracted water (c. 13°C) provides convincing evidence that water was being drawn from the hypolimnion at the time.

The preceeding discussion provides some evidence for the likely influence of the subsurface offtake structures, in Warragmaba Dam, on horizontal movements within the hypolimnion of Lake Burragorang. If these effects were to be confirmed then it is also reasonable to expect some vertical effect, particularly from the operation of the HEPS offtake, and also to predict that the rapid changes involved in opening and closing of the offtake may affect the patterns of harmonic motions of the thermocline (cf. Wunderlich 1971).

Finally, it should be noted that the discussion has focussed only on years in which the complicating consideration of inflows can be largely ignored for the periods under consideration, at least in respect of "effective inflows" as determined from the Inflow Register (see Inflow Register; Chapter 3, Table 3.3). Nevertheless, it is not possible to dismiss the effect of inflow altogether, if only because the HEPS is operated only when the lake is very nearly full.

CHAPTER 5

A COMPARISON OF THERMAL AND OXYGEN STRATIFICATION IN WET AND DRY YEARS

INTRODUCTION

The extremely important role of inflow as a modifier of Lake Burragorang's thermal and oxygen stratification behaviour is indicated in previous chapters. Inflows have a role in restricting vertical circulation and disrupting the normally holomictic pattern of the thermal cycle in the lake, and they apparently also influence the vertical temperature structure during the relatively extended period of stratification. In both instances, there is an indeterminate effect of the subsurface offtake structures at Warragamba Dam, but one that is certainly significant.

A general effect of inflow is to promote the comparatively variable behaviour found in this lake, and this is in full agreement with Allanson's (1973) summation of the physical limnology in man-made lakes in which he expresses the view (cf. Straskraba 1973; same volume) that residence time and density currents are the two factors that differentiate between the limnology of reservoirs and lakes. Baxter (1977) reiterates this view, and notes the extra complexity of behaviour that arises from subsurface discharge. Allanson (1973) also comments that the effects of inflow into natural lakes, particularly small lakes, may have been underestimated.

In this context, I believe there is a clear advantage in the study of reservoirs in relation to inflow effects, in that measurement of water balance is usually fundamental to reservoir management, so that detailed records of inflow and outflow are often available. For Lake Burragorang, the combination of this data with long-term records of some physical, chemical and biological data enables a useful assessment of the interaction between inflows and the important cycles in the lake (ie temperature and oxygen stratification), permitting a relatively great selectivity in the data to be analysed. Several other factors make Lake Burragorang of particular interest.

The very great range of retention times recorded for the lake (c. 0.5 - 22 years; Lake volume/annual total inflow minus evaporation) provides a particularly broad base for the study of inflow into a single basin. Also the non-seasonality of inflow means that the study encompasses the interaction of inflow with the full annual cycle of the lakes behaviour.

In the present treatment, a quantitative comparison is made between four "Wet" years and four "Dry" years (a total of 40% of the available data), again using monthly means. An account of the analytical method, with some general commentary on the analysis and the years chosen for the analysis, is given in the Methods section (Chapter 2). Some of this information is briefly repeated here.

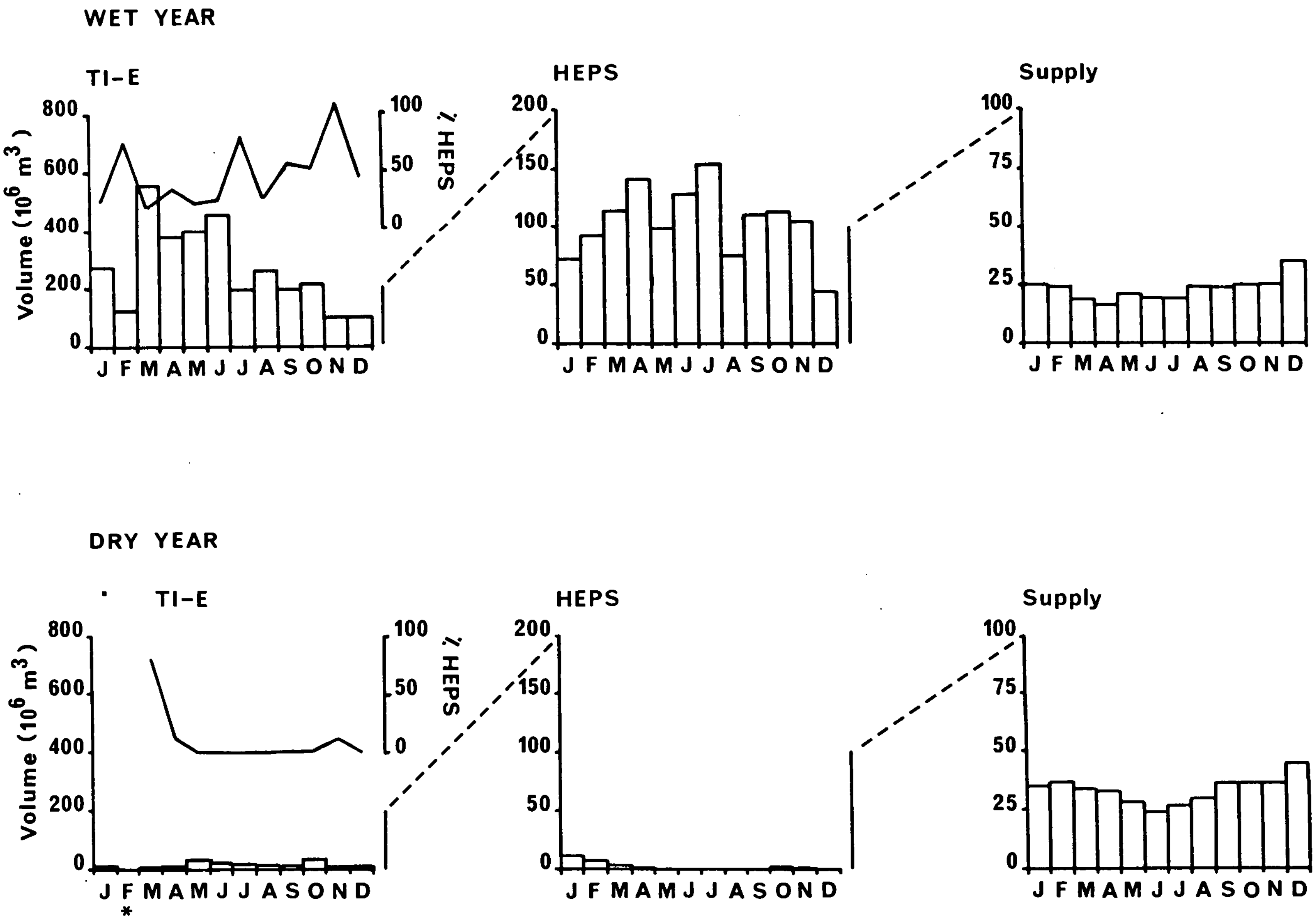
The wet years (1963, 1974, 1976, 1978) had total annual inflows (minus evaporation) ranging from 2074 - 3904 10^6 m^3, yielding retention times less than 1 year (0.53 - 0.99 years). The dry years (1965, 1968, 1979, 1980) had inflows totalling 259 10^6 m^3, or less, giving retention times from 7.9 - 22.8 years. The comparison is, therefore, between data blocks that differ in annual inflow (minus evaporation) by about an order of magnitude. These two groups of years will be referred to as WET and DRY to differentiate, in this and subsequent sections, between specific reference to this analytical experiment and general discussion of wet and dry years, albeit extrapolated from the results of this analysis.

Fig. 5.1 shows the monthly inflows for WET and DRY years, averaged over the four years, and the mean monthly outflow volumes for both the Hydro-electric (HEPS) and water supply (to Prospect Reservoir) offtakes. The HEPS volumes, expressed as a percentage of the inflow volume, are also given. It should be noted that the choice of years was governed by the attempt to include years that were consistently wet or dry, rather than having aspects of both. From Fig. 5.1 it is evident that, although there is greater inflow for every month of the WET years than for corresponding DRY months, there are a

FIGURE 5.1

Four year means of monthly total inflow minus evaporation (TI-E), outflow through the HEPS (centred at 44.5 m below FSL), and the outflow to Prospect Reservoir (Supply; variable offtake c. 0 - 60 m), for WET (1963, 1974, 1976, 1978) and DRY (1965, 1968, 1979, 1980) years. A line, representing the HEPS outflow volume expressed as a percentage of the total inflow is plotted on the inflow graph.

* In February the mean of total inflow minus evaporation was negative, so no value is given for the % HEPS. Consequently, the January % HEPS appears as a single point, separated from the remainder of the line.



WET YEAR
TI-E
HEPS
Supply
Volume (10^6 m^3)
% HEPS
800
600
400
200
0
100
50
0
200
150
100
50
0
100
75
50
25
0
J F M A M J J A S O N D
DRY YEAR
TI-E
HEPS
Supply
*

few WET months with relatively low inflow averages, particularly February, November, and December. Consequently, these months may be expected not to reflect the direct influence of inflows to the same extent as some of the other months, such as March - June.

Nevertheless, the statistical comparison of mean monthly profiles (temperature (°C) and dissolved oxygen (mg l^{-1}), taken from consistent depth intervals (relative to the lake surface), provides a means of quantifying the differences in Lake Burragorang's thermal structure and oxygen cycle resulting from inflow and the operation of the Warragamba Dam offtake structures. Unfortunately, it is difficult to adequately differentiate between inflow and outflow effects, given the direct correlation between inflows and the operation of the mid-depth hydro electric offtake at the dam (a function of management policy). To this extent the results of this comparison are not directly comparable with the response of a natural lake to inflows, despite the probable overlap of retention times considered here with those of some natural lakes.

Theory

For the purpose of this analysis, it may be assumed that the DRY year group represents the closest analogy between Lake Burragorang and a natural lake. In this condition, Lake Burragorang's thermal and oxygen cycles are expected to be governed by the interaction of factors familiar to classical limnology (ie the predominance of wind forced circulation, and oxygen demand resulting from the decay of in situ biological production). The DRY years, then, are a control against which the effects of inflow and outflow can be measured.

The effects of inflows on the thermal profile will depend on whether interflow or underflow occurs, but the following discussion provides a conceptual framework for the analysis:-

If it can be assumed that turbulent mixing associated with the passage of an inflow into a lake is either non-existent, as has been assumed in early thermal models (ie the WRE ~~(~~1969~~)~~ model reported in Park and Schmidt 1973), or significant at the point of entry into the lake (Slotta 1973), as has been assumed in a more recent study (Imberger and Hebbert 1980), then the effect an inflow has at a point further down the lake depends solely on the displacement of a volume of water eqivalent to the volume of the inflow (plus that entrained at the point of entry, in the latter case). In order to satisfy the requirement for conservation of mass an inflow must cause upward displacement of all the water above its level of entry (Carmack and Farmer 1982), and an outflow must result in downward movement of all water overlying the depth of the offtake (assuming selective withdrawal from a layer horizontally projected at the level of the offtake; Park and Schmidt 1973). If this simplified conceptual framework is accepted, then the effect of advection on the thermal profile, at site 3D, can be expressed as follows:-

An interflow, penetrating above the HEPS offtake (ie in the period from about October/November through to May/June) will tend to displace warmer, overlying water upwards and result in heat loss by flooding over the dam-crest. Counter to this, however, is the effect of the HEPS which results in a bulk downward movement of the overlying water by subtracting a deeper layer, and enhances the downward movement of heat. The HEPS may actively promote heat storage in the lake under these circumstances. This is described for Fontana Reservoir by Wunderlich (1971).

An inflow penetrating to site 3D below the level of the HEPS offtake (an underflow, June - October approximately) will displace all the water in the lake upwards, cooling the lake by causing surface outflow. The effect of the HEPS will be to ameliorate this cooling, by withdrawing water at least a little cooler than the surface layers that would otherwise be spilled from the lake.

It is important to note the close relationship between inflow and mid-depth outflow in determining the thermal profile measured relative to the surface. It is probably a reasonable assumption that the mid-depth outflow, by itself, would have little effect on the vertical distribution of heat relative to the surface, unless it drew the water level down to the extent of subtracting metalimnetic water. Some change might result from re-distributing the volume of warmed water deeper into the lake basin (an effect of changing volume distribution with depth).

Of course the assumption that turbulence is not associated with the passage of an inflow through a lake, is certainly an over-simplification. In the preceeding account, the downward vertical movement of heat is regarded as a response to sub-surface outflow, with the role of inflow confined to the insertion of a metalimnetic layer above the level of the offtake; that is, the interflow does not directly contribute to the downward movement of heat. But the limnological textbook literature, although not exlicit on this point, certainly depicts a broadening of the metalimnion in response to metalimnetic inflow (see Hutchinson 1957; Wetzel 1983), and this appears to involve the downward movement of heat. Also, there are indications in the literature that inflow enhances the downward movement of heat in natural lakes with surface outflow. Weiss et al (1979) found a threefold increase of the apparent vertical diffusivity (cm^2 sec^{-1}) in Lake Constance (Bodensee), in the wetter of two years which differed in yearly outflow by about 20%.

It seems, therefore, that the effects of inflow and outflow on the thermal structure in the lake can be qualitatively similar, which emphasizes the difficulty of making any meaningful separation of these effects, given that the two processes occur together in a lake primarily used for water supply and situated where inflows can be unreliable, making the operation of the high volume HEPS offtake a luxury mostly confined to times of flood.

The following account is organised into descriptive sections according to the following headings, heat content, stability, Birgean wind work, temperature (Absolute means, and differences between monthly means), and dissolved oxygen (absolute, and rate comparison) followed by an overview of the differences between WET and DRY year groups in which the general effects of inflow and outflow are discussed. Differences between the two groups are expressed as a percentage of the DRY year value, unless specifically stated to the contrary.

HEAT CONTENT AND MECHANICAL STABILITY IN WET AND DRY YEARS

Heat Content (cal cm^{-2})

The annual progression of heat content is shown in Fig. 5.2, and the absolute and percentage differences between WET and DRY years are given in Table 5.1. Because of the mode of calculation (use of a single monthly profile for each group rather than making a separate heat content calculation for each of the four years), there is no direct indication of the statistical significance of the differences between the two groups. Nevertheless, there is a clear trend, in WET years, towards a greater heat content during the first four months of the year. This advective heat gain is sufficient to maintain an increasing lake heat content until March, in WET years, and sustain almost the same heat content in April. In contrast, the DRY year heat content peaks in February, and has fallen by 5% of this value in April, when the two groups achieve their maximum divergence of 5900 cal cm^{-2} (12%). In Fig. 5.3, the approximate advective heat exchange for WET years is shown, based on the monthly means of inflow/outflow volumes and temperatures, and the mean temperature profile at site 3D. I have assumed that advected heat is insignificant in the DRY year, and can be ignored. The figure indicates that heat is gained by the lake during the first three months of the year, and this gain (expressed per unit lake area), ranging from c. 6000 - 7000 cal cm^{-2}, is

FIGURE 5.2

Annual progression of heat content (cal cm^{-2}; x 4.19 = Joules cm^{-2}), Schmidt stability (gm-cm cm^{-2}; x 9.807 10^{-15} = Joules cm^{-2}), Birgean wind work (gm-cm cm^{-2}), and volume weighted mean dissolved oxygen concentration (mg l^{-1}). Birgean wind work has been calculated relative to an initial condiion of circulation at 4°C. The actual wind work is, therefore, calculated relative to the minimum winter value (cf. Heide 1982), in a similar manner to the calculation of Birgean annual heat budget. In the present case, the data must be assumed to be circular, so that the January - April values of wind work can be calculated relative to the August minimum.

Symbols:-

WET years (+)

DRY years (·)

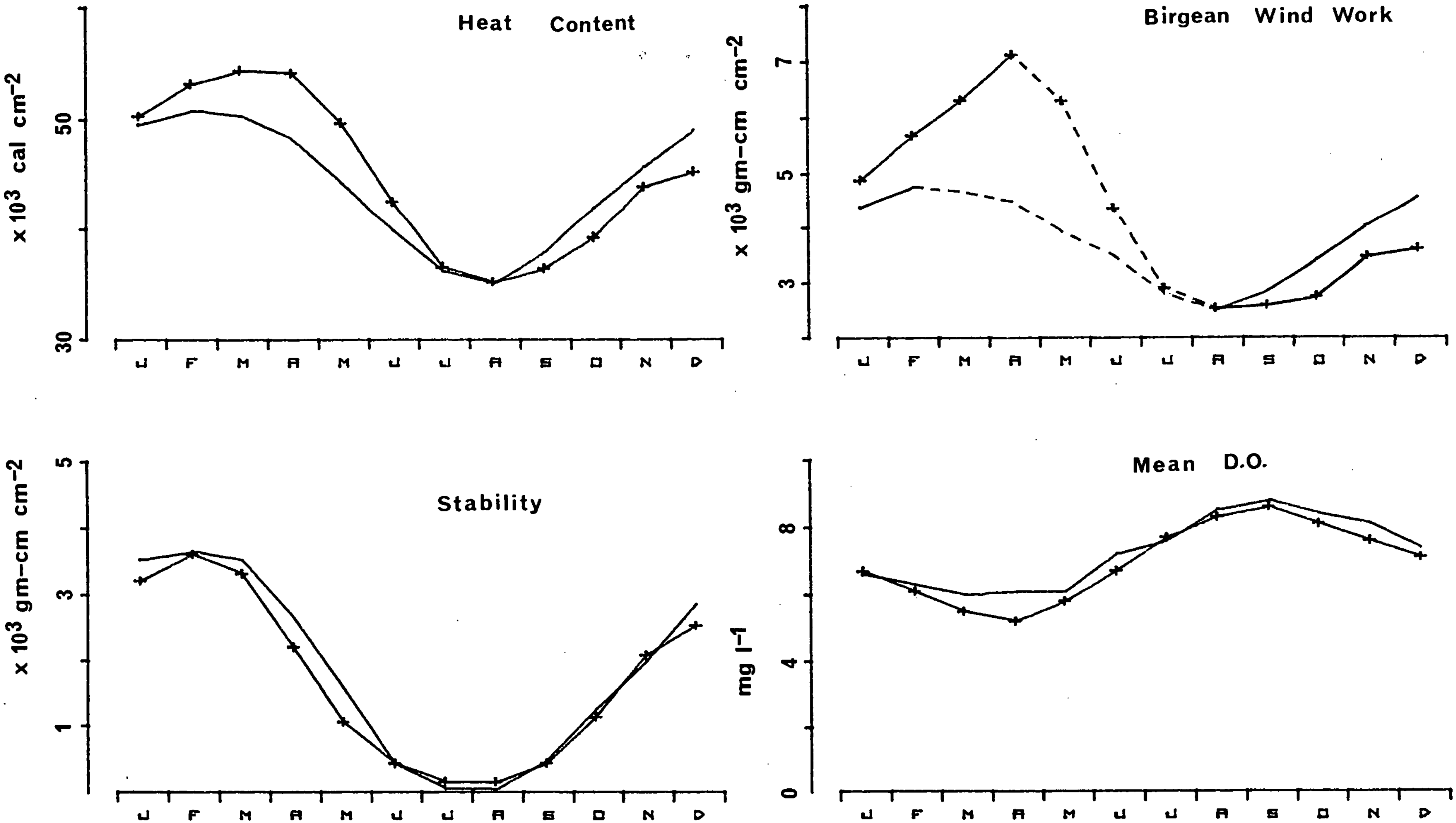
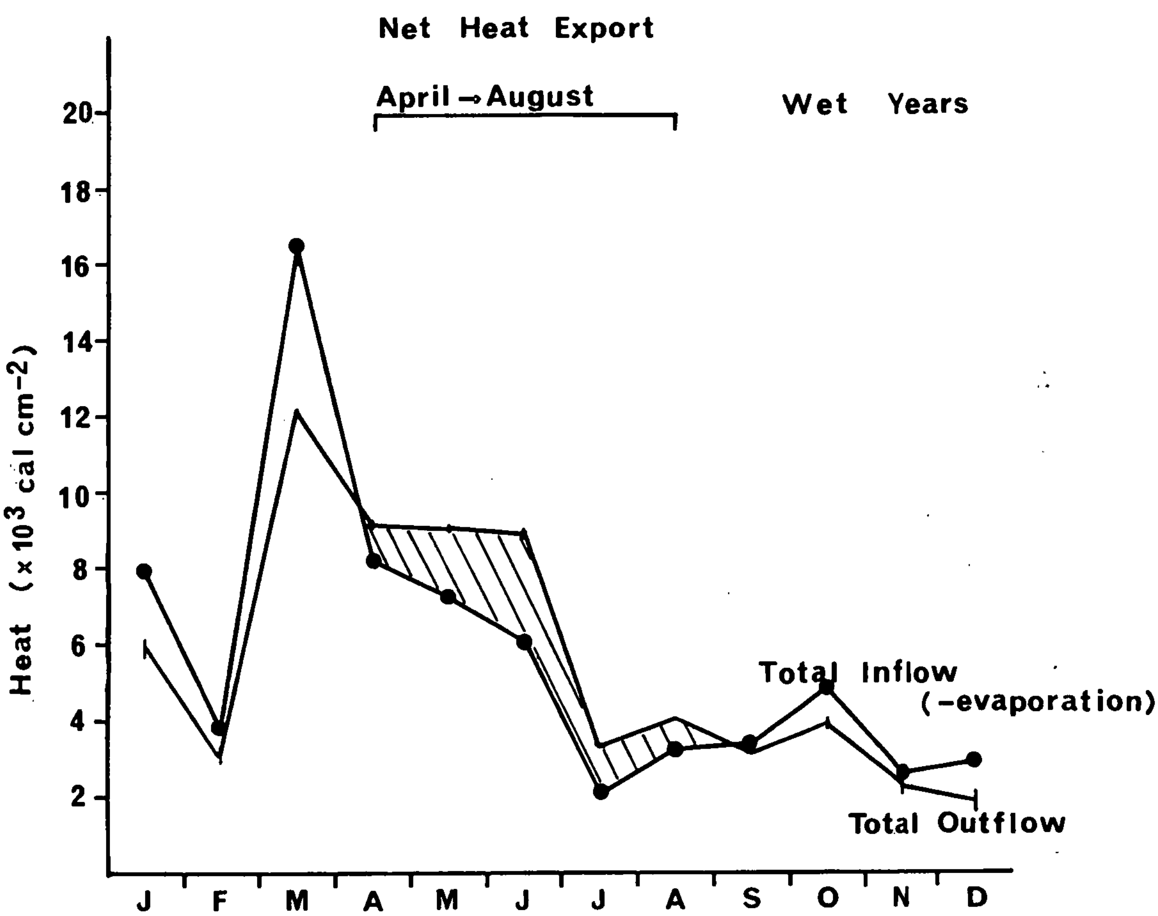

Heat Content
x 10^3 cal cm^-2
50
30
J F M A M J J A S O N D
Birgean Wind Work
x 10^3 gm-cm cm^-2
7
5
3
J F M A M J J A S O N D
Stability
x 10^3 gm-cm cm^-2
5
3
1
J F M A M J J A S O N D
Mean D.O.
mg l^-1
8
4
0
J F M A M J J A S O N D

Table 5.1 Absolute, and percentage differences between WET and DRY year groups. Heat content, Idso-modified Schmidt stability, and Birgean wind-work. Calculated as DRY minus WET, and expressed as a percentage of the DRY year value.

	DRY - WET Heat Content (cal cm^{-2})		DRY - WET Stability (gm-cm cm^{-2})		DRY - WET Birgean wind-work (gm-cm cm^{-2})		DRY - WET Ave. Oxygen (mg l^{-1})	
	Difference	Percent	Difference	Percent	Difference	Percent	Difference	Percent
JAN.	-730	1.5	320	9.1	-440	23.2	-0.1	1.5
FEB.	-2380	4.7	40	1.1	-890	39.4	0.2	3.2
MAR.	-3990	7.9	210	6.0			0.5	8.3
APR.	-5860	12.1	460	17.3			0.9	14.8
MAY	-5320	12.0	520	32.9			0.3	4.9
JUN.	-2660	6.7	-10	2.3			0.5	6.9
JUL.	-330	0.9	-100	167			-0.1	1.3
AUG.	-110	0.3	-110	275	0	0	0.2	2.4
SEP.	1430	3.8	40	8.3	280	82.4	0.2	2.3
OCT.	2820	6.7	110	8.8	710	77.2	0.3	3.6
NOV.	1730	3.8	-100	5.1	580	38.2	0.5	6.2
DEC.	3700	7.6	320	11.2	980	47.6	0.3	4.1

FIGURE 5.3

Approximate advective heat exchange, for Lake Burragorang, during WET years. Four year means of inflow temperatures, and thermal profiles (site 3D; see Fig. 5.4) have been used in the calculation. Heat input and output are expressed per unit surface area of the lake at FSL (75 km^2), so as to be directly comparable with the heat content calculations. Because the water supply offtake (to Prospect Reservoir) operates between 0 m and c. 60 m, the heat export from this offtake has been calculated using both surface and bottom temperatures at site 3D. Consequently the heat content of the total outflow is in fact a range, and this is depicted with vertical bars in the figure. It is apparent that for WET years the uncertainty introduced by this technique is insignificant.



Net Heat Export
April → August
Wet Years
Heat (x 10^3 cal cm^-2)
20
18
16
14
12
10
8
6
4
2
Total Inflow
(-evaporation)
Total Outflow
J
F
M
A
M
J
J
A
S
O
N
D

at least sufficient to account for the maximum difference in whole lake heat content between WET and DRY years.

From April to July both WET and DRY years show a near linear decline in heat content, reaching very nearly the same minimum heat content in August (Fig. 5.2). This requires a greater loss of heat during the WET years than is recorded for the DRY years, and the rates of decline in heat content (from Fig. 5.2) are -200 cal cm^{-2} day^{-1} and -132 cal cm^{-2} day^{-1} respectively; a 50% faster rate of cooling for the WET years. Fig. 5.3 shows a nett loss of heat from the lake due to the combined effects of inflow and outflow (c. 6400 - 7300; April - July), which is sufficient to explain the enhanced rate of decline in heat content for the WET years, by approximately balancing the advective gains of the first few months of the year. Both WET and DRY groups record a very similar annual minimum heat content in August (c. 35200 cal cm^{-2}), differing by less than 1% (Table 5.1).

With the onset of heating, in September, the relative positions of the two groups is reversed (ie DRY > WET), and the DRY year heat content remains higher for the remainder of the year (Fig. 5.2). In September the WET year heat content is c. 4% (1400 cal cm^{-2}) less than that of the DRY year group, and Fig. 5.2. gives the impression of a delayed beginning to the spring heating phase, for the WET years. The advective heat balance, for August, represents a net export of heat of c. 600 - 1100 cal cm^{-2}, while in September the influx and efflux of heat are virtually balanced (Fig. 5.3). Thus the effects of the inflow/outflow relations may again be invoked to explain the difference between the WET and DRY years, if these advective effects are applied to the following months heat content. This lagged comparison was also used earlier (ie comparing the April maximum difference in WET - DRY heat content with the summed advective effect of January - March, for WET years), and may be justified on two grounds. First, there is a certain delay between the initiation of an inflow, and the time when it has its maximum effect at site 3D. This may be a matter of only a few days, in many instances, but comparison of the daily records of lake level with the chemical profiles taken at site 3D, indicates that periods of weeks may be involved in some cases. Second, and possibly more important, is the fact that detection of an inflow and its effects at site 3D is a function of the approximately two weekly sampling regime, and this may increase the apparent delay between inflow initiation and the recorded time of its maximum effect. Nearly half of the registered inflows have a period of 14 days or more between the initiation of inflow and the date of maximum recorded turbidity effect at site 3D. In an analysis based on monthly averages, it is therefore possible that an apparent lag of one month may arise between an effect measured in the catchment and the time it becomes important at site 3D.

Another plausible explanation for the delayed start to WET year heating is that of climate, as it is conceivable that WET years are characterised by generally less solar irradiance than DRY years, and that surface heating is weaker in this early part of the heating period.

Wet year heat content remains 4 - 8% lower than that of the DRY years for the remaining months of the comparison (Table 5.1), despite the fact that the lake apparently gains heat (c. 1300 - 1500 cal cm^{-2}) from September to November of WET years (Fig. 5.3); this may be significant and indicate that the two groups differ climatically, it certainly indicates that in the absence of WET year advective heat gain the two groups might have diverged even further.

It is interesting to compare the December and January heat contents for the two groups, to determine whether there are heating or cooling trends relative to the initial heat contents for WET and DRY years. It is reasonable to assert that the DRY years are part of a warming trend in the lake, while the WET years represent a counterbalancing cooling trend. This is certainly in keeping with the observed trends in hypolimnetic temperature in relation to successive wet or dry years (cf. Chapter 3, Fig. 3.1; temperature isopleths), although the hypolimnion contains only a small percentage of the overall lake volume. This apparent divergence of heat content during the course of a year, may have implications for the comparison of WET and DRY year heat budgets, also. Assuming circularity of the data, these are 15700 cal cm^{-2} yr^{-1} (DRY) and 19100 cal cm^{-2} yr^{-1} (WET), an increase of c. 22% in WET years. However, in successive WET years, the retarded spring heating period may decrease the maximum heat content of the following year and reduce the annual heat budget.

Schmidt Stability (gm-cm cm^{-2})

Fig. 5.2 contains the annual progression of Idso-modified Schmidt stability (gm-cm cm^{-2}) for the WET and DRY years, and numeric and percentage differences are given in Table 5.1. Unlike the heat content, there is no change in the timing of maximum or minimum stability as a result of inflow/outflow effects. The annual maximum stability (c. 3600 gm-cm cm^{-2}) occurs in February, and WET and DRY year groups differ by only 1%, although in January and March the WET years are from 6 - 9% less stable. Both groups show a fairly rapid decline of stability between March and June. At its maximum, the rate of this decline is the same for each group (-37 gm-cm cm^{-2} day^{-1}; expressed as a daily rate, and assuming one month equivalent to 30.4 days), but whereas the WET year decline is maximal between March and May and slows between May and June, the DRY years achieve their maximum rate of decline between April and June. The maximum absolute difference between WET and DRY years is in May, when the WET year group is c. 500 gm-cm cm^{-2} (Table 5.1), or 33%, less stable than the DRY year group. This reflects the combined influence of metalimnetic inflow and mid-hypolimnetic outflow, which cause the metalimnion to be "spread out" and to become less steep.

Much the same temperature difference between epilimnion and hypolimnion, occupies a broader depth zone in WET years, compared to the DRY years.

May and June, are the months in which the transition to underflow occurs, and the effect of cold underflows entering the hypolimnion, is to slow the rate of decline of the WET year stability, relative to that of the DRY years. In June, both groups have stabilities of c. 400 gm-cm cm^{-2}, but in July and August the WET year group (150 gm-cm cm^{-2}; Fig. 5.2) is up to 3 times as stable as the DRY year group (c. 50 gm-cm cm^{-2}; Fig. 5.2). Minimum stability occurs in August, and neither group exhibits a very high winter stability, but it is equally true that neither group records zero stability in the course of the year, on the basis of these mean monthly profiles. The absolute difference in August stability (c. 100 gm-cm cm^{-2}), shows a measurable effect of cold underflows in stabilising the water column during WET years.

From September to December, the WET year stability is usually from 9% - 11% less than that of the DRY year group, with the exception of November, in which the WET year stability is a little greater than for DRY years. The absolute differences between the two groups are of the order of 50 - 100 gm-cm cm^{-2}, until December when the WET year stability is c. 300 gm-cm cm^{-2} less than for DRY years. During the heating phase both groups show a near linear increase in stability from September to December, of about 24 gm-cm cm^{-2} day^{-1}.

Birgean Wind Work (gm-cm cm^{-2})

The annual progression of Birgean wind work is also given in Fig. 5.2. The cooling phase is marked as a broken line, as it is unsatisfactory to view the work of the wind decreasing simply because the lake is cooling. Birgean wind work, as has been mentioned previously (see Materials and Methods, Chapter 2), is conceptually unable to cope with advective effects. This was also noted by Johnson and Merritt (1979), because Birgean wind work is calculated relative to an assumed historic starting point, it has a conceptual time factor. Also, it regards all heating as being a surface generated phenomenon. Schmidt stability, on the other hand, escapes these limitations by being conceptually instantaneous, with no dependence on the history of the stratification. The reason for including the Birgean wind work here, is to assess the advective effect, and to re-assess its importance with respect to the long term mean data discussed in Chapter 3.

In January Birgean wind work is c. 23% greater for the WET year group than for the DRY years (Table 5.1), and the two groups continue to diverge until April, when the WET year maximum occurs. In DRY years Birgean wind work peaks in February. Comparison of the two maxima, the wind work required to distribute the summer heat income, clearly shows the effect of advected heat as the WET year value (c. 4600 gm-cm cm^{-2}; April) exceeds that of the DRY years (c. 2300 gm-cm cm^{-2}; February) by 100%. This demonstrates the sensitivity of the Birgean wind work calculation to inflow and outflow effects, which have substantially changed both the magnitude and timing of the annual maximum. These changes are similar, though more extreme, than the advective effects on the lake heat content.

The winter (August) minimum values of Birgean wind work (c. 2500 gm-cm cm^{-2}, calculated relative to a circulating water column at 4°C) are used here as the initial values (0 gm-cm cm^{-2} of wind work) for calculation of wind work in this warm monomictic lake (following Heide 1982). Heating begins in September, and between September and December the DRY year Birgean wind work increases to about 2100 gm-cm cm^{-2} in a near linear progression, while the WET year Birgean wind work increases slowly between August and October, and is only about 1100 gm-cm cm^{-2} in December (48% of the DRY year value). Again the influence of advection is quite marked.

The particular sensitivity of the Birgean wind work calculation to advective effects, requires some re-interpretation of the heating efficiency index (maximum Birgean wind work/summer heat income; gm-cm cal^{-1}). It would have been reasonable to expect that the heating efficiency of WET years would appear greater than that of the DRY years, given that advection results in deeper penetration of heat into the water column, and increases the heat content during WET years. However, the advective effect on Birgean wind work is such that the ratio is greater (indicating less efficiency) for the WET years. It therefore seems likely that the actual ratio for Lake Burragorang, in its least advectively modified form, is c. 0.14, which still denotes reasonably inefficient heating, but lies within the range of values reported by Hutchinson (1957) for natural lakes of the world.

THERMAL STRATIFICATION IN WET AND DRY YEARS

Mean Temperature Profiles

Fig. 5.4 shows the profiles of WET and DRY year temperatures. Each of the points in the profiles represents a mean (n= 4) and the two groups have been compared with a t-test (see Materials and Methods, Chapter 2). Significant differences between the profiles are marked with groups of stars (p = 0.05 *, p = 0.01 **, p = 0.001***, p = 0.0001****).

In January, there are no significant differences between the two groups, although the WET year profile shows a slightly lower epilimnetic temperature and higher hypolimnetic temperature, on average (Fig. 5.4). The metalimnetic gradient is also slightly less steep. The first appearance of statistically significant differences is in February, when the WET year profile is significantly warmer at 24 m and 30 m (p = 0.05; Fig. 5.4). In the period from February to May, the metalimnetic region is affected by advective processes, such that the WET years become significantly warmer in the region between c. 18 m and 48 m below the lake surface (Fig. 5.4). The deepest affected layer is 60 m, in May. Differences, significant at p = 0.001 and p = 0.0001, occur

FIGURE 5.4

The figure shows four year mean temperature profiles (°C; site 3D) for WET and DRY years. The method of calculating the means is detailed in Materials and Methods (Chapter 2), and is the same as that used to calculate the 20 year mean profiles, except that n = c. 4 rather than c. 20. the profiles have been compared using a t-test (see Steel and Torrie 1981; p 106) which does not assume equivalent variance. The mean temperature for each depth has been compared in this manner, and significant differences between the profiles are marked to the right of each plot. The last plot compares December and January, assuming that the data is circular. This assumption is not strictly valid for averages derived from four non-contiguous years.

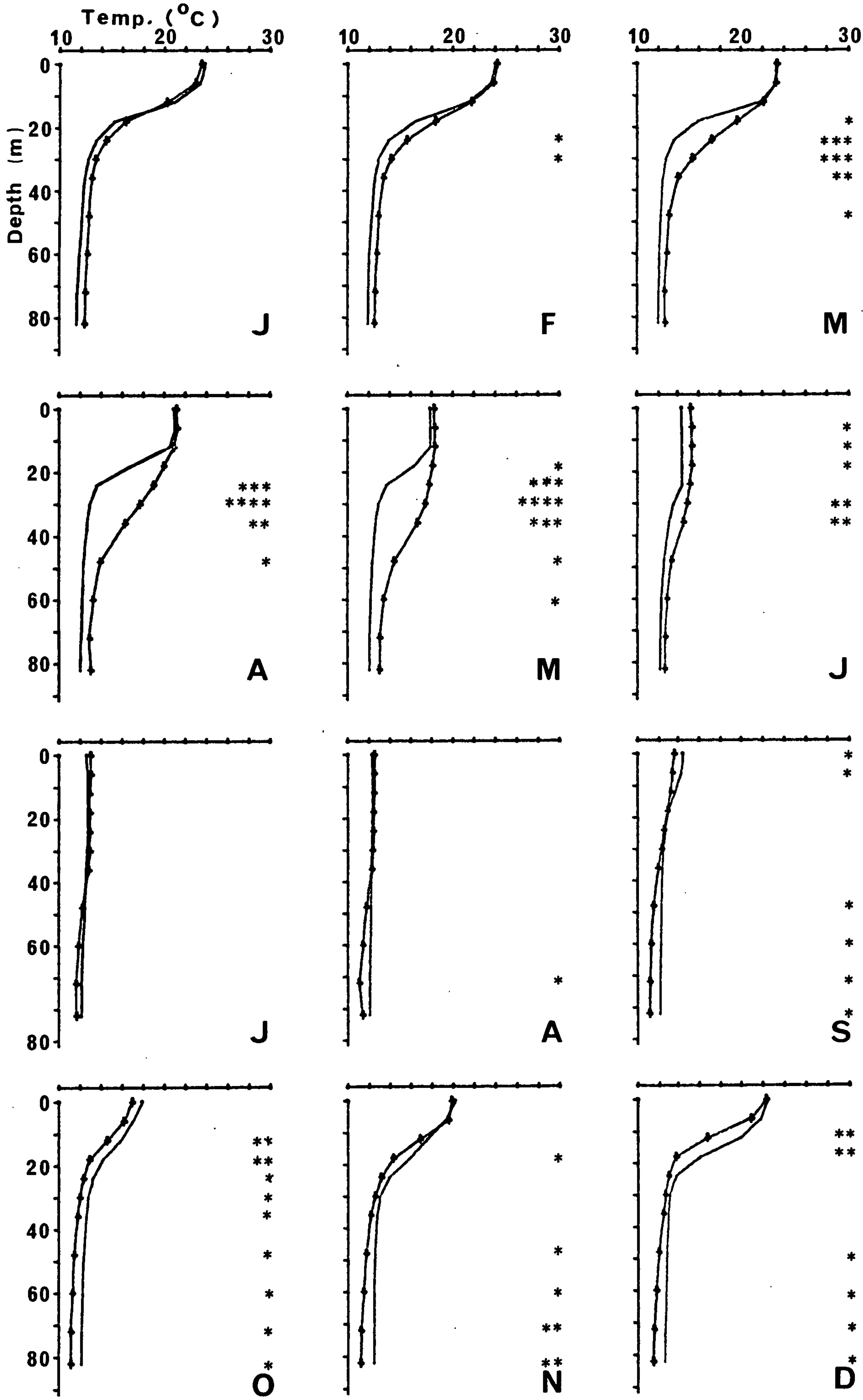
Sigificance levels are denoted by the number of stars:-

none	=	not significant
*	=	significant at 5%, p = 0.05
**	=	significant at 1%, p = 0.01
***	=	significant at 0.1%, p = 0.001
****	=	significant at 0.01%, p = 0.0001

Symbols:-

WET years (+)

DRY years (·)



Temp. (°C)
10
20
30
Depth (m)
0
20
40
60
80
J
F
M
A
M
J
J
A
S
O
N
D

between 24 m and 36 m in March, April and May. The combined effect of inflow and outflow, in the early part of the year, is to promote the downward movement of heat into the water column, broadening the WET year metalimnion and reducing the steepness of the thermal gradient. This is consistent with the increased heat content of the WET years, in this period. A comparison of the depth of the thermocline (plane of maximum rate of change of temperature with depth) between the two groups, shows that in May the WET year thermocline lies between 36 m and 48 m (nominally 42 m; taking the mid-point between sample depths), while the DRY year thermocline is between 18 m and 24 m (nominally 21 m), some 20 m shallower. Interestingly, the depth of the convectively mixed epilimnion, appears similar for each of the groups with the possible exception of May, when the WET year epilimnion appears to penetrate deeper into the water column (perhaps 12 - 18 m deeper). It will be seen from consideration of the data for dissolved oxygen that this impression is deceptive, and that the two groups have similar mixed depths at this time (within the accuracy of the profiling). Throughout these first five months of the year, the temperature of the near surface and deepest layers of the water column do not differ significantly between the two groups, although, as previously mentioned, the WET year hypolimnon is slightly warmer.

That the transition from interflow to underflow, occurs in May and June, is demonstrated by the distribution of effective inflows (from the Inflow Register; Chapter 3, Table 3.3) for the WET years, in Table 5.2. This change removes the direct effect of inflow from the metalimnion to the hypolimnion, and results in active cooling of the WET year hypolimnion. This can be seen from the June plot of mean temperature profiles (Fig. 5.4), as a narrowing of the gap between the WET and DRY year hypolimnetic temperatures. The pattern of WET to DRY year difference changes in June, with significant differences extending from about 6 m to 36 m. This is the culmination of a trend of

Table 5.2 Distribution of 16 registered inflows, occurring in the four WET years (1963, 1974, 1976, 1978). The distribution is first given, according to the dates upon which the flows were initiated, and then according to the sample date of maximum turbidity effect at site 3D.
I, Interflow. U, Underflow. Subscripts denote the year of the inflow, more than one inflow may occur within a single month.

	Inflow initiation date	Total	Date of Maximum Turbidity	Total
JAN.				
FEB.	I_{76}	1		
MAR.	I_{74}: I_{78}	2	I_{74}: I_{76}: I_{78}	3
APR.	I_{63}: I_{74}: I_{78}	3	I_{78}	1
MAY	U_{74}	1	I_{63}: I_{74}	2
JUN.	I_{63}: U_{63}: I_{78}: U_{78}	4	I_{63}: U_{74}: I_{78}	3
JUL.	U_{63}	1	U_{63}: U_{63}: U_{78}	3
AUG.	U_{63}: U_{74}	2		
SEP.			U_{63}: U_{74}	2
OCT.	U_{76}	1		
NOV.			U_{76}	1
DEC.	I_{63}	1	I_{63}	1

slightly slower surface heat loss in WET years, which appears in Fig. 5.4 as a slowly widening gap between the epilimnetic temperatures of the two groups from April to June. The statistical significance, or otherwise, of this trend is better evaluated from the month to month difference plots in the following section.

The zone of most significant difference in June ($p = 0.01$), is from 30 - 36 m, below the DRY year thermocline (nominally 27 m), and above that of the WET year group, which remains between 36 m and 48 m (nominally 42 m). Interestingly, this latter depth zone is within the region directly affected by the HEPS withdrawal structure. The significance of the temperature difference is declining in June (Fig. 5.4).

In July and August, the effect of underflows is to further cool the WET year hypolimnion relative to the deep layer temperature of the DRY year group. In this period, when convective circulation is at its peak, there are essentially no significant temperature differences between the two groups. The difference, at 72 m in August, is based on a WET year temperature mean from data for 1963 and 1978 only, compared to the full four years of data for the DRY year value; regardless of the significance ($p = 0.05$), this comparison does not fulfill the intent of the analysis. The WET year profile retains a thermocline, at a nominal depth of 42 m in both months, concluding a four month period in which this discontinuity is maintained in the region directly affected by the HEPS withdrawal current. Though not shown in Fig. 5.4, the WET years do not achieve full vertical overlap of standard deviation bars in either July or August. In comparison, the DRY year thermocline, which begins a pronounced downward movement in June, is weakly present in July (this is supported by the oxygen data, which will be described presently). The gradient is diffuse, however, probably indicating considerable vertical variability in the position of the thermocline for individual years. The standard deviations overlap vertically for the whole water column in July, which is taken to indicate the possibility of isothermy at this time for the DRY year group, and is consistent with the descriptive account of the thermal stratification cycle (see Chapter 3). The DRY year profile is approximately isothermal in August, having a total gradient of 0.3°C, which is generally monotonic; though it should be noted that a temperature discontinuity of only 0.4°C (between 48 m and 61 m) was associated with an oxycline of 5.3 mg l^{-1} (50% of saturation) on the 5th of July 1966, demonstrating the potential stability of deep thermoclines in Lake Burragorang.

Surface heating begins in September, for both groups, but the DRY year epilimnion is significantly warmer ($p = 0.05$, 0 - 6 m; Fig. 5.4) in the September profile. Deep water cooling also continues between August and September of WET years, and the region from 48 m to the deepest sampled depth is significantly cooler ($p = 0.05$; Fig. 5.4) than for the DRY years. The WET year profile lacks any distinct thermocline in September, while the DRY year thermocline is re-developing between about 6 m and 18 m below the surface. For the remainder of the year, the profiles develop more or less in step, with the hypolimnia remaining different at the 5% probability level ($p = 0.05$) at least. By November, the WET year surface temperature is virtually the same as that of the DRY year group. A region of difference does develop below the epilimnion in December, with the WET year metalimnion overlying that of the DRY years. This seems at odds with the anticipated effect of metalimnetic inflows that are expected to occur from about November. However, Table 5.2 indicates that the effective inflows (September - December) were underflows, in September, October, and November, with a single interflow recorded for December. It is difficult to explain how an advective effect could lead to the situation recorded for December (Fig. 5.4), given that both interflow and the HEPS outflow are established as promoting a weaker metalimnetic thermal gradient and enhancing the downward movement of heat. It should be remembered that advective influences are weaker in this period (November - December) than at other times for this comparison. A possible explanation for the apparently shallower zone of heating in WET years, is that this group is characterised by a relatively turbid epilimnion, following the winter period of maximum vertical circulation. Mean Secchi depth ranges from 2.0 m (September) to 3.2 m (December) for WET years, compared to 4.7 m (September) to 5.4 m (November; 4.5 m in December) for DRY years. In a comparison of two small turbid (Secchi depths < 1 m) reservoirs, Scheibe et al (1975) found that the more turbid reservoir developed a shallower epilimnion and had a lower heat content, effects they attributed to the greater albedo of the more turbid water. A similar situation may develop in Lake Burragorang during WET years.

<u>Temperature change from month to month</u>

Differences between successive months of the WET and DRY group profiles (effectively rates of temperature change month^{-1} for each depth) are shown, in Fig. 5.5. These have also been tested (t-test) for significant difference between the WET and DRY year groups, at each depth (see Materials and Methods, Chapter 2). It should be noted that the month to month differences have not been tested for significant variation from zero, so that any discussion of positive or negative temperature change, for WET or DRY years individually, represents a largely qualitative assessment. A standard deviation was determined for each point, so that it is possible, where necessary, to provide some statistical quantification. Because it is cumbersome to make continual reference to significant differences between month to month difference plots, the latter will be abbreviated to DP (ie the February - March DP) in this section.

The January - February DP (Fig. 5.5) shows no significant difference between the WET and DRY year groups, although the WET year temperature changes are slightly greater (c. 0.5°C), in the upper 30 m of the water column,

FIGURE 5.5

The figure shows four year mean profiles of month to month temperature differences (°C $month^{-1}$; site 3D) for WET and DRY years. The method of calculating the means is detailed in Materials and Methods (Chapter 2), and is the same as that used to calculate the 20 year mean difference profiles, except that n = c. 4 rather than c. 20. the profiles have been compared using a t-test (see Steel and Torrie 1981; p 106) which does not assume equivalent variance. Significant differences between the profiles are marked to the right of each plot. Zero change in temperature, between successive monthly mean profiles of temperature (°C) is marked with vertical line in each of the plots. The significance or otherwise of the changes are for a comparison between the two lines, and not for a comparison between each profile and the zero line. The last plot compares December and January, assuming that the data is circular. This assumption is not strictly valid for averages derived from four non-contiguous years.

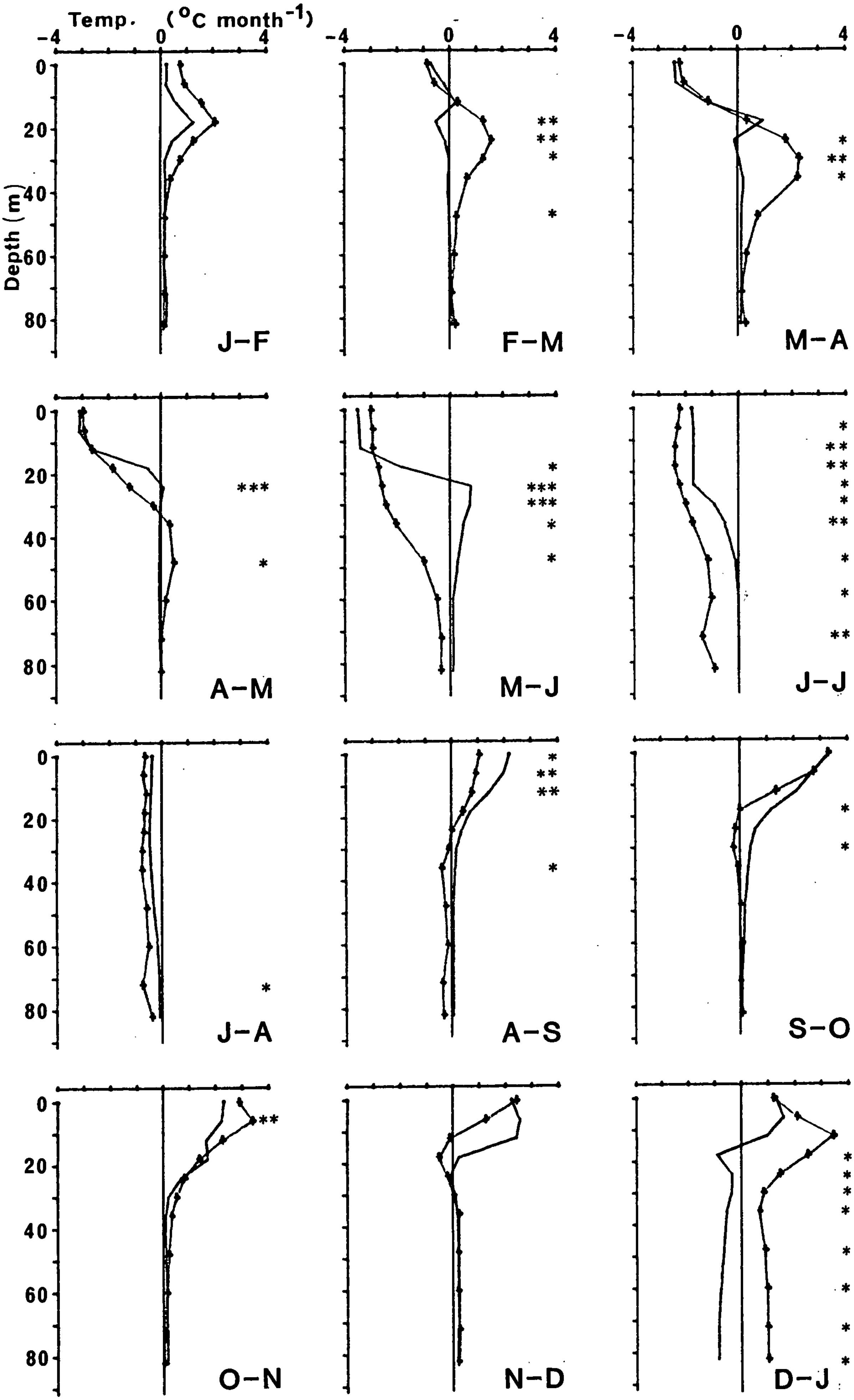
Sigificance levels are denoted by the number of stars:-

none	=	not significant
*	=	significant at 5%, p = 0.05
**	=	significant at 1%, p = 0.01
***	=	significant at 0.1%, p = 0.001
****	=	significant at 0.01%, p = 0.0001.

Symbols:-

WET years (+)

DRY years (·)



Temp. (°C month⁻¹)
-4
0
4
Depth (m)
0
20
40
60
80
J–F
F–M
M–A
A–M
M–J
J–J
J–A
A–S
S–O
O–N
N–D
D–J

than for the DRY years. The shape of the curves is similar, with both showing slight surface heating and having a region of active heat gain centered on the 18 m sample depth, indicating a rate of downward heat transfer in the metalimnion (cf Fig. 5.4), in excess of the rate at which heat is being absorbed at the lake surface. This was discussed previously in relation to the 20 yr. mean monthly difference plots (see Chapter 3), and marks the later stages of the heating phase when an epilimnion (detectable from the 6 m sample intervals) first becomes a feature of the thermal profile in the lake.

In the period between February and May, significant differences emerge as the form of the WET year plots change in response to the effects of advective processes. Differences are recorded in the region between 18 m and 48 m (p = 0.05 - 0.001; Fig. 5.5), associated with wave-like zone of heating in WET years, which has its peak at 24 m in the February - March DP, moving down to 48 m in the April - May DP. This contrasts with the DRY year plots, which give some indication of a very restricted region of heat gain at 12 m in the February - March DP, which moves to 24 m in the April - May DP (Fig. 5.5). In both cases the points are within 1 standard deviation of zero, which also applies to the regions of slight negative temperature change immediately beneath them. This uncertainty is understandable, given the vertical dynamism of the thermocline. I consider that this highlights the difference between DRY years, in which the depth of the thermocline is more or less stable or changing only slowly at this time (February to April), and WET years, in which there is a clear and dominant downward heating trend caused by advection. The upper water column of both groups loses heat in this period, and the maximum negative differences between successive months is similar. In the April - May DP, the WET years are losing heat from deeper in the water column than the DRY years.

In the May - June DP the zone of significant difference still occupies the 18 m - 48 m depth zone, with greatest significance at 24 m and 30 m (p =

0.001). The DRY year plot shows the only definite downward heating wave for the cooling season, having its peak in the region 24 m - 30 m beneath the surface, and possibly extending down to c. 48 m (Fig. 5.5; though this point is within 1 standard deviation of zero). This represents the period when the DRY year epilimnion is beginning to deepen relatively quickly. Prior to this, the DRY year epilimnion remained within 6 m - 12 m of the surface, showing a tendency to deepen slowly from 6 m or less in January to include the 12 m sample depth by May (Fig. 5.4). The WET year group, on the other hand, is losing heat at all depths in the May - June DP, which shows the first effects of underflows on the WET year thermal profile. The WET year differences are more than 1 standard deviation from zero at all depths down to 48 m, and this comparison (May - June) probably represents the greatest behavioural deviation between the two groups of years. This difference relates primarily to the depth from which heat is lost, and the fact that the DRY year group is only just beginning to show an obvious deepening of the mixed zone, with the accompanying downward propagation of heat stored in the epilimnion during the summer heating phase, while the WET year (advectively influenced) group has entered a phase of heat loss that affects (though not evenly) the whole water column. Remembering, that while in DRY (convection dominated) years, loss of heat at any depth (ie a negative temperature change in the DP) essentially requires mixing to that level, an inflow or outflow affects all layers above it, so that a fully negative DP (May - June; Fig. 5.5) results from a cold underflow at this time of year. The rate of surface cooling (between March and June) is a little less for WET years than for DRY years (Fig. 5.5), though this difference is not statistically significant. This may appear to contradict the fact that the WET year heat content is known to be declining more rapidly than the DRY year heat content, in this period. However, the depth and therefore volume from which heat is being lost, is greater for the WET year group, and this counterbalances the more rapid but superficial DRY year temperature decline.

Between June and August the water column loses heat in both cases, but for DRY years the region below c. 50 m is little changed, while for WET years there is marked cooling of the deeper layers, especially between June and July (Fig. 5.5) when there is significant difference from 6 m - 60 m (the 72 m comparison is not accepted for either the June - July or July - August DP's). There is no significant difference between the two plots of the July - August DP (ignoring 72 m comparison), but the WET year plot is slightly more negative and this heat loss does not taper off, in the deeper layers, to the extent seen for the DRY year plot (Fig. 5.5).

Surface heating is evident from the August - September DP, and is the major feature for the remainder of the year. The August - September DP shows a significantly lesser rate of heating, for WET years, in the upper 12 m of the water column. At the same time, for the WET year group, the region below c. 30 m is still cooling as a result of cold influent water. In the period from September to December the WET year temperature gain, at the surface, is at least as great as that of the DRY years, but for WET years the zone of active heating is shallower in the September - October DP (this is associated with significant differences at 18 m and 30 m) and the November - December DP (not significant, as the DRY year differences at 6 m and 12 m have quite high standard deviations). The only other significant difference is recorded for the October - November DP, when the WET year heat gain, at 6 m, is greater than that for the DRY years.

Assuming an hypothetical circularity, to permit the calculation of December - January differences, is interesting in that it clearly shows that relative to the January starting point (when there are no significant differences; Fig. 5.4) the WET year profile shows a definite cooling below about 18 m, while the DRY year profile warms; the two groups show significant differences ($p = 0.05$) from 18 m to the deepest sampled depth in the December - January DP. A similar conclusion was reached from examination of the lake heat content. In the long-term, then, it is likely that Lake Burragorang's hypolimnetic temperature, which ranges from 10.5 - 13.7°C, based on monthly means of deepest sample from 1961 - 1980 (ignoring the records, > 15°C, from 1961 - 1962), drifts up and down between these limits, according to the pattern of inflows. Dry years show a gradual warming trend, which is interrupted by rapid declines associated with the cold underflows of wetter years. In certain circumstances, advective effects act to promote temperature increase in the hypolimnion, most notably the 1961 flood, but no such catastrophic effect has been recorded since. A slightly enhanced warming in the hypolimnion, can be observed during the WET year cooling phase (March - May; this is best observed in Figs 5.8 and 5.9, which will be presented in the overview section), but the cooling effect of underflows is much greater.

OXYGEN STRATIFICATION IN WET AND DRY YEARS

Mean Dissolved Oxygen Profiles

The four year mean oxygen concentration profiles (mg l^{-1}) for WET and DRY year groups are plotted in Fig. 5.6. It will be observed from Fig. 5.6 that the distance between the plots of WET and DRY oxygen concentration which constitutes a significant difference, is large relative to that typical of the previous temperature comparisons, and that it varies with the time of year, being greatest in the period just before hypolimnetic re-oxygenation and least in the early stages of stratification (c. October). This reflects the comparatively great variation that develops, particularly in the hypolimnetic oxygen concentration, during the long period of thermal stratification, and results in large standard deviations of the 4 year means for each depth.

The two groups begin the year with virtually identical dissolved oxygen

FIGURE 5.6

Four year mean profiles of dissolved oxygen (mg l^{-1}; site 3D) for WET and DRY years. The method of calculating the means is detailed in Materials and Methods (Chapter 2), and is the same as that used to calculate the 20 year mean profiles, except that n = c. 4 rather than c. 20. the profiles have been compared using a t-test (see Steel and Torrie 1981; p 106) which does not assume equivalent variance. The mean dissolved oxygen for each depth has been compared in this manner, and significant differences between the profiles are marked to the right of each plot. The last plot compares December and January, assuming that the data circular. This assumption is not strictly valid for averages derived from four non-contiguous years.

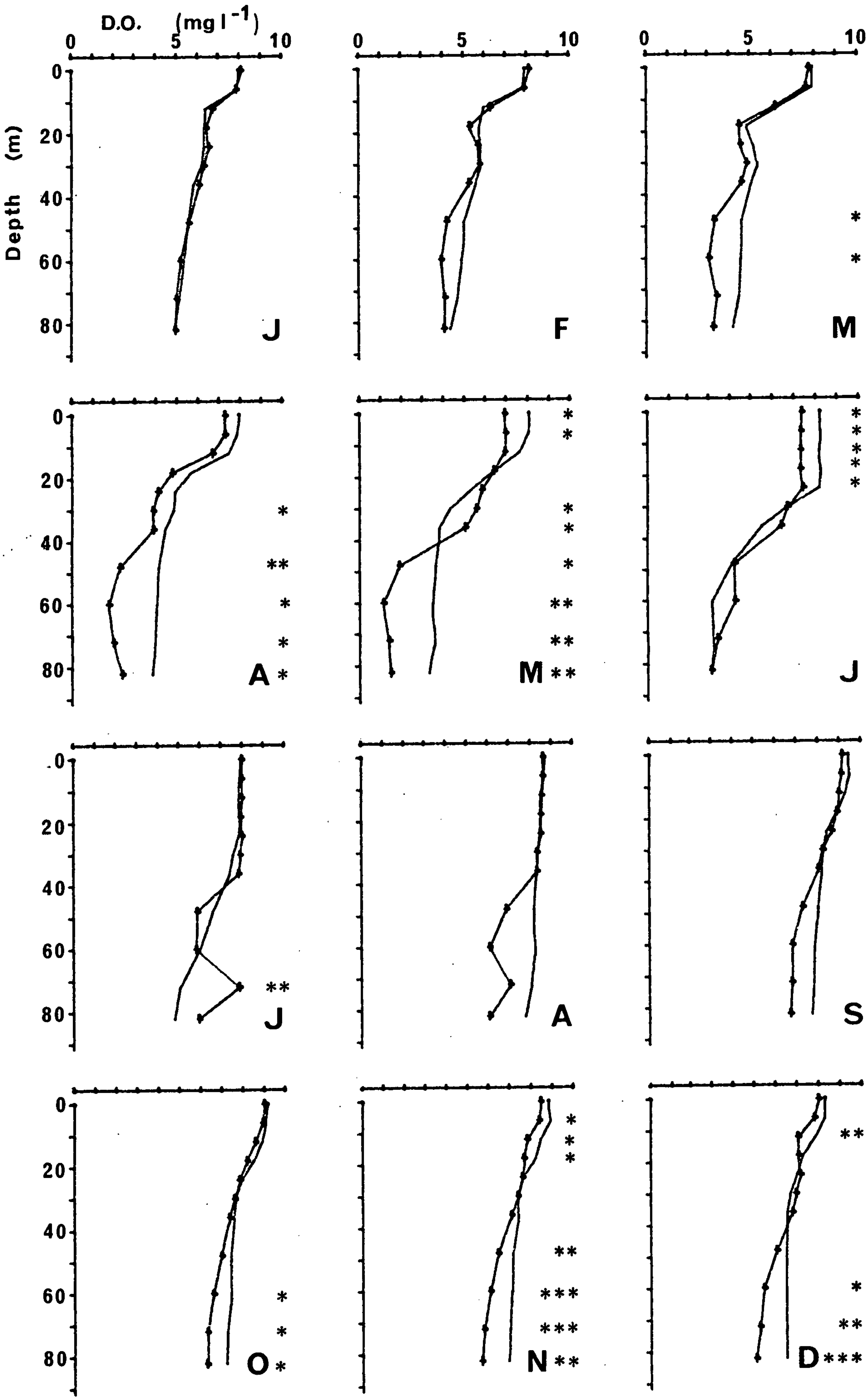
Sigificance levels are denoted by the number of stars:-

none	=	not significant
*	=	significant at 5%, p = 0.05
**	=	significant at 1%, p = 0.01
***	=	significant at 0.1%, p = 0.001
****	=	significant at 0.01%, p = 0.0001

Symbols:-

WET years (+)

DRY years (·)


D.O. (mg l -1)
0
5
10
Depth (m)
0
20
40
60
80
J
F
M
A
M
J
J
A
S
O
N
D
*
**
***

profiles (January; Fig. 5.6). Epilimnetic concentration is c. 8 mg l^{-1}, the oxycline resides between 6 m and 12 m (nominally 9 m), and the hypolimnetic concentration has a gradient from c. 6.5 - 5 mg l^{-1}. It is not until March that significant differences arise between the WET and DRY profiles. In February both groups show a negative heterograde profile, and there appears to be a tendency for the WET year group to show both a lower hypolimnetic oxygen concentration, and for the minimum concentration to occur at mid-depth in the hypolimnion (Fig. 5.6). In March, the WET year hypolimnion has significantly lower oxygen concentration than is found for the DRY year group, at 48 m and 60 m ($p = 0.05$; Fig. 5.6). The epilimnetic concentrations are much the same (c. 7.8 mg l^{-1}), and the metalimnion lies between 6 m and 18 m in each case. The development of the metalimnetic trough (negative heterograde oxygen profile) is at its peak for both groups, in March, but is a little more pronounced in the WET year profile, though not to the point of statistical significance.

In April and May, as the epilimnion begins to gradually deepen, the WET year epilimnetic oxygen concentration declines in relation to that of the DRY years, and is about 1.1 mg l^{-1} (11% of saturation) less in May (significant at $p = 0.05$; 0 m - 6 m; Fig. 5.6). In WET years, this lower epilimnetic oxygen concentration presumably develops in response to a greater oxygen demand, associated with suspended particulate matter brought in by metalimnetic inflows. Hypolimnetic oxygen concentrations, also differ significantly ($p = 0.05 - 0.01$; Fig. 5.6) between the two groups in April and May, particularly in the region below c. 50 m, where the mean concentrations differ by about 2 mg l^{-1} (c. 3.5 mg l^{-1}, 33% saturated, DRY, compared to 1.5 mg l^{-1}, 14% saturated, WET).

In this first part of the year, when the metalimnion is directly affected by inflowing water, the WET and DRY year profiles develop differently, in that the WET year profile comes to have essentially two oxyclines. This may also be said of the DRY year profile during those months when there is a metalimnetic dissolved oxygen trough, but to a lesser degree, and for a shorter period. The WET year profile has a distinct oxycline between 36 m and 48 m (nominally 42 m) in February, which is maintained, at the same depth, throughout this period (Fig. 5.6). In April and May the shape of the WET year profile is consistent with the effect that might be generated by large metalimnial inflows; a zone of similar oxygen concentration bordered by oxyclines at the upper and lower boundaries. It is also notable, however, that the deeper oxycline forms in the region adjacent to the HEPS offtake (c. 35 m - 54 m). Interestingly, the depth of the epilimnion remains the same for both groups for these first five months, with the possible exception of May, in which the WET year epilimnion appears to be fully circulating to 12 m, while for DRY years the oxygen concentration begins to decline beneath the 6 m sample depth. This latter observation is much less obvious in the comparison of temperature profiles (see Fig. 5.4).

In June, the epilimnion (0 m - 24 m for both groups) is the only zone of significant difference ($p = 0.05$; WET < DRY; Fig. 5.6), as the WET year hypolimnion is now directly affected by cold, oxygenated underflows, and the four year means (for each depth) are quite variable. For example, the oxygen concentration of the deepest samples, in June (3.1 mg l^{-1}, WET and DRY; Fig. 5.6), have standard deviations of 3.2 mg l^{-1} (WET) and 0.9 mg l^{-1} (DRY). In DRY years oxygen continues to be depleted in the zone below 60 m, and June is the month of maximum oxygen gradient. The concentration declines from 8.2 mg l^{-1} (80% of saturation) in the epilimnion, to a minimum of 3.2 mg l^{-1} (29% of saturation) in the hypolimnion, a total gradient of 5 mg l^{-1} (51% of saturation). The maximum gradient for the WET year group occurs in May, prior to the beginning of the underflow period, and is from an epilimnetic concentration of 7.0 mg l^{-1} (74% of saturation) to about 1.4 mg l^{-1} (13% of saturation) in the hypolimnion, a total gradient of 5.6 mg l^{-1} (61% of saturation). The combined effect of inflow and outflow is, therefore, to increase the oxygen demand, initially in the metalimnion and hyplimnion, but affecting the whole water column in the autumnal period of increased convective mixing. A comparison of the mean oxygen concentrations for WET and DRY year groups (calculated using LIMNO/2, and taking into account the changing volume distribution with depth) is given in Table 5.1, and the seasonal progression of mean dissolved oxygen concentration is plotted in Fig. 5.2. This shows a maximum difference of 0.9 mg l^{-1} in April (WET < DRY, by 15% of the DRY year average oxygen concentration).

There are no significant differences between WET and DRY year groups, from July to September (ignoring the data for WET years, at 72 m, in July and August). This is indicative of the great variability of the WET year hypolimnetic concentrations, in particular. In this period, however, there do appear to be some qualitative differences between the two groups. The DRY year group shows the changes appropriate to convective overturn, ending in the approximately monotonic gradient of the August profile. This contrasts with the WET year group, which appears to maintain some vertical discontinuity in the oxygen profile, centred on the region adjacent to the HEPS offtake. The comparatively great variability of the WET year data, and the unfortunate lack of some data for the WET year profiles in July and August, can be seen from the representative monthly profiles in Appendix 2. Consequently, the preceding observations are tentative. In DRY years, oxygen stratification is established in September, mainly by the continued absorption of atmospheric oxygen by the near surface layer. A more pronounced stratification is apparent for the WET years, but the variability of the WET year hypolimnetic concentration ensures that the two groups do not differ significantly.

Following the re-establishment of thermal and oxygen stratification, the hypolimnion is the first region in which significant differences emerge. The

WET year group has significantly lower oxygen concentration ($p = 0.05 - 0.001$) below about 60 m, for the remainder of the year (Fig. 5.6). In November and December there are also significant differences in the upper water column, with the WET year group showing a significantly lower oxygen concentration ($p = 0.05$; Fig. 5.6) between 6 m and 18 m in November. In December, the WET year profile has a slight metalimnetic trough (not apparent in the DRY year profile) and has a significantly lower oxygen concentration at 12 m ($p = 0.01$; Fig. 5.6). In November and December, the WET year profile again tends to develop a secondary oxycline between 36 m and 48 m. I consider that the persistence of this cline, at a specific depth for much of the year, is related to the effects of the HEPS withdrawal current at site 3D, irrespective of the equivocal nature of the data for the winter months.

<u>Changes in Dissolved Oxygen Concentration from month to month</u>

The difference plots (DP), describing changes in oxygen concentration between the monthly profiles of Fig. 5.6, are presented in Fig. 5.7. Significant differences between the WET and DRY group DP's are marked on each plot.

A significantly greater rate of volumetric hypolimnetic oxygen depletion is evident for the WET years between January and April, and probably in the April - May DP also (Fig. 5.7). The zone of significant difference varies, but is generally 48 m or deeper. In the March - April and April - May DP's the epilimnion is also a region of significant difference ($p = 0.05$; 0 m and 6 m), with the DRY years showing little change while the WET year surface layers decline in oxygen concentration. The WET and DRY group curves are of similar shape in this part of the year (January - April), though the WET year curves are consistently to the left of the DRY year curves, indicating a more negative, or less positive, change in oxygen concentration between months. Both groups show a tendency to develop a metalimnetic trough, while the WET year trend towards a mid-depth hypolimnetic trough contrasts clearly with

FIGURE 5.7

The figure shows four year mean profiles of month to month differences in dissolved oxygen concentration (mg l^{-1} month^{-1}; site 3D) for WET and DRY years (The method of calculating the means is detailed in Materials and Methods, Chapter 2). Significant differences between the profiles are marked to the right of each plot. Zero change in dissolved oxygen, between successive monthly mean profiles is marked with vertical line in each of the plots. The two lines have been compared for statistical significance at each depth. The differences have not been test for significance relative to the zero change line. The last plot compares December and January, assuming that the data is circular. This assumption is not strictly valid for averages derived from four non-contiguous years.

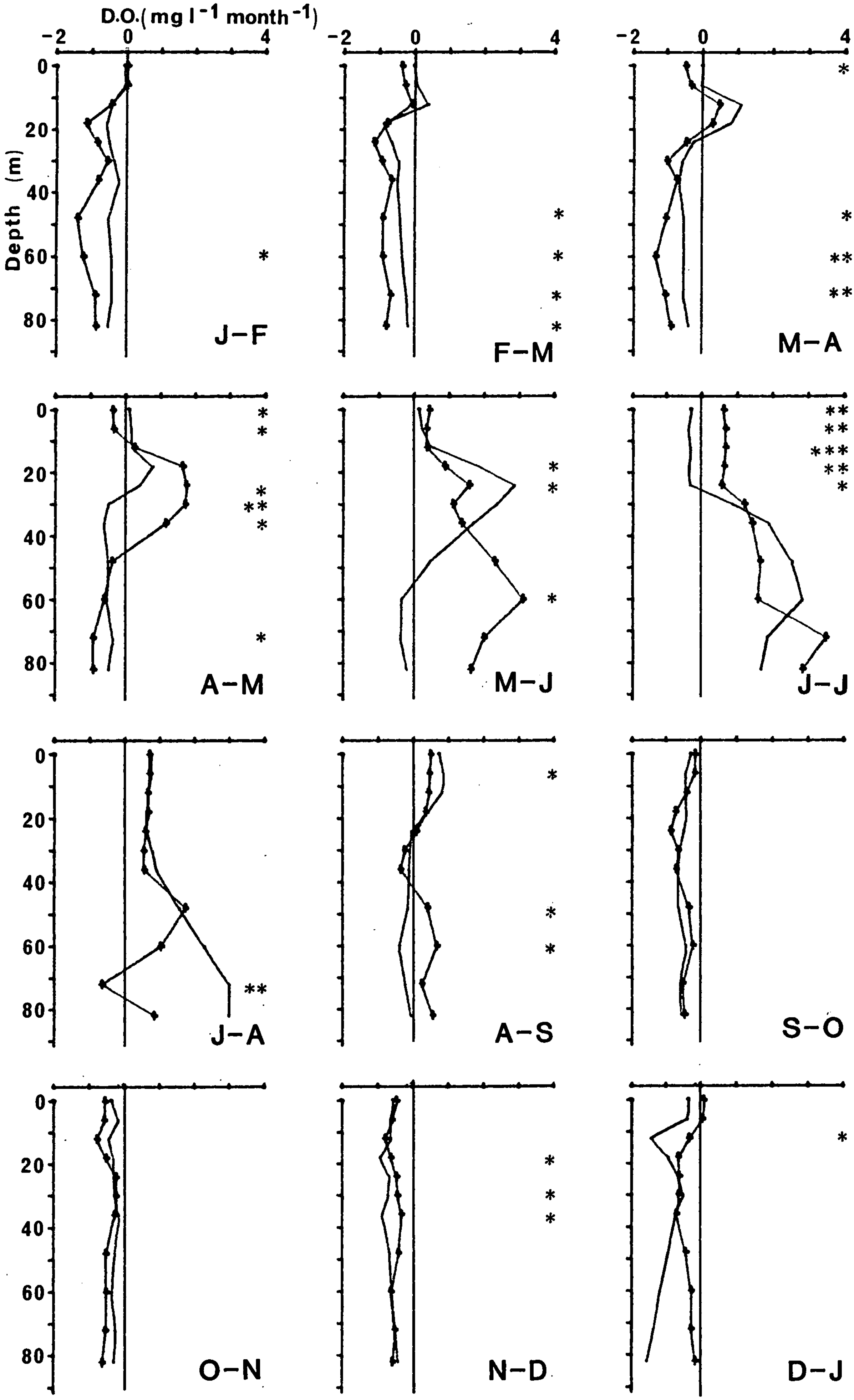
Sigificance levels are denoted by the number of stars:-

none	=	not significant
*	=	significant at 5%, p = 0.05
**	=	significant at 1%, p = 0.01
***	=	significant at 0.1%, p = 0.001
****	=	significant at 0.01%, p = 0.0001.

Symbols:-

WET years (+)

DRY years (·)


D.O. (mg l-1 month-1)
-2
0
4
Depth (m)
0
20
40
60
80
J–F
F–M
M–A
A–M
M–J
J–J
J–A
A–S
S–O
O–N
N–D
D–J

changes in the DRY year hypolimnion (Fig. 5.7).

The dominant feature of the difference plots in the autumn and early winter, is the wave-like zone of increasing oxygen concentration, that, for a purely convective system, accompanies the deepening penetration of convective mixing in the lake. The behaviour of the two groups differs considerably in this period. In DRY years, this zone is most obvious from March to June, with the maximum positive change (crest of the wave) at c. 12 m in the March - April DP (Fig. 5.7), moving down to c. 24 m in the May - June DP. For WET years, this zone of increasing oxygen concentration, is more pronounced than that of the DRY years in the April - May DP, both in terms of the rate of positive change, and the vertical extent of the affected region; the two curves differ significantly between 24 m and 36 m ($p = 0.05 - 0.01$; Fig. 5.7). The greater rate of dissolved oxygen increase, in WET years, is probably the result of metalimnetic inflows bringing in well oxygenated water from the catchment, while the enhanced downward penetration is presumably attributable to the operation of the HEPS offtake, and perhaps also to turbulence associated with the inflows.

In the May - June DP the WET years gain in dissolved oxygen concentration throughout the water column, showing the first effects of deep interflows (c. 40 m - 60 m) and underflows. However, with the variability of the deep water oxygen concentrations in the WET years, the only significant difference occurs at 60 m ($p = 0.05$; Fig. 5.7). The DRY years continue to lose oxygen in the hypolimnion, but in the metalimnion there is a region of significantly greater oxygen increase compared to the WET years ($p = 0.05$; 18 m - 24 m; Fig. 5.7). This results from the first major deepening of the DRY year group's mixed depth for the year. The WET years show a smaller peak at 24 m, which is also caused by the deepening of the mixed layer. It will be remembered that the two groups have similar convectively mixed depths in May and June (Fig. 5.6).

In the June - July DP, the two groups are significantly different from 0 m - 24 m (p = 0.05 - 0.001; Fig. 5.7). The changes in this zone have opposite signs, with the WET years gaining oxygen, while the DRY year concentration declines. The increase, in the WET year mixed layer oxygen concentration, balances the losses of the period from February to May, and the two groups have similar near surface concentrations in July (Fig. 5.6). There are no significant differences below 24 m, and the WET years continue to actively increase in oxygen concentration as a result of inflows, while, in the DRY years, the downward wave of oxygenation has its peak at 60 m, but affects the water column to the deepest sampled depth (Fig. 5.7). This deepening progression, of DRY year re-oxygenation, culminates in the July - August DP when the maximum positive change in oxygen concentration is at the deepest sampled depth. Ignoring the 72 m comparison, there are no significant differences until the August - September DP (p = 0.05, at 6 m, 48 m and 60 m; Fig. 5.7). Both groups increase in oxygen concentration near the surface, but the DRY year increase exceeds that of the WET year group. In the hypolimnion the two groups show opposing trends, with the WET year concentration still actively increasing in response to underflows, while the newly isolated DRY year hypolimnion has begun to stagnate.

There are no further significant differences between the two groups until the November - December DP, in which the DRY year change is significantly more negative (p = 0.05) in the region from about 18 m to 36 m, below the surface (Fig. 5.7). This difference coincides with the trend towards development of a second oxycline in WET years, at the level of the HEPS offtake (c. 42 m). Between September and December, both groups decline in oxygen concentration throughout the water column.

If the difference between December and January is considered (ie assuming circularity) the DRY year curve is to the left of the WET year curve, indicating that the DRY years have gained in oxygen concentration to a greater extent than the WET years. This is especially so in the deeper layers of the hypolimnion, where the WET years show little change, and in the metalimnion. Overall, it appears that there is a trend towards increasing oxygen concentration throughout the water column in dry years, although the differences are within one standard deviation of zero from 0 m to 6 m.

COMPARING OXYGEN AND THERMAL CYCLES

It is interesting to compare the temperature and dissolved oxygen difference plots (Fig. 5.5 and Fig. 5.7) during the cooling phase that preceeds the winter period of maximum vertical circulation, as this illustrates the effect of advection on the downward propagation of heat and dissolved oxygen. Despite the fact that many of the points (particularly the month to month differences of dissolved oxygen) lie within 1 standard deviation of zero, this need not invalidate observations that rely, primarily, on the relative shapes of the plotted curves.

The feature of most interest concerns the relationship between the downward penetrating zones of heat and dissolved oxygen for the two groups, particularly in the period from February to June. Direct comparison of Fig. 5.5 and Fig. 5.7, shows that for DRY years the peaks of increasing temperature and oxygen (crests of the descending waves) occur at similar depths, throughout this period. This is most obvious in the May - June DP, when the DRY year mixed layer deepens markedly, but in the preceeding month the small heating peaks are definitely within the regions of oxygen increase (Figs 5.5 and 5.7). In contrast, the WET year heating wave is consistently deeper than the corresponding region of increasing dissolved oxygen, with the peaks vertically separated by about 20 m, in the March - April and April - May DP's (Figs 5.5 and 5.7). This comparison underlines the extra behavioural complexity, introduced by inflows in combination with subsurface outflows.

In DRY (convection dominated) years, during the cooling period, temperature and dissolved oxygen progress more or less in step, with changes in dissolved oxygen dependent on the changes in the mixed depth. Changes in dissolved oxygen concentration are, therefore, to a great extent predictable from a knowledge of the thermal behaviour. In WET years, however, the contribution of advection is to uncouple this relationship to some extent. This accentuates the differences in behaviour that arise from the fact that temperature is essentially conserved except at the surface, while dissolved oxygen is continually being depleted throughout the water coulmn. In a convective system it may be reasonable to know the thermal behaviour, and rely on that to predict attendant phenomena like dissolved oxygen distribution, but in a system where advection is significant this is less realistic.

A further point of interest, is that this analysis supports the conclusion that inflow and outflow do not (within the resolution of this analysis; 6 m vertically) affect the convectively mixed layer in the cooling period, though this might appear to be the case from examination of the temperature data alone. This should be compared with Welsh (1984) who definitely concludes that the high level offtake, in Dartmouth Reservoir, increases the mixed layer depth.

SUMMARY/OVERVIEW OF THE WET AND DRY YEAR COMPARISON

The results of the analytical experiment presented in this chapter, demonstrate a significant effect of advective processes on the stratification of temperature and dissolved oxygen in Lake Burragorang (at site 3D). Advection affects both the vertical structure, and the seasonal progression of temperature and dissolved oxygen profiles measured at this site (adjacent to Warragamba Dam). By extension of this data, advection also markedly changes the whole lake parameters of heat content, Schmidt stability and Birgean wind work.

In this analysis of WET and DRY years, the annual retention times of the four DRY years are sufficiently long that the lake is considered to behave as a system dominated by convective mixing, while the four WET years measure the change in stratification behaviour arising from a roughly ten-fold increase of annual inflow, which brings the annual retention time to less than 1 year. Straskraba (1973), suggests that the assumptions, upon which models of thermal stratification in reservoirs are based, must be changed when retention time falls below a critical value, somewhat less than 1 year.

To assist in summarising the present chapter, Figs 5.8, 5.9, 5.10, and 5.11 show the annual progression of temperature (Fig. 5.8, DRY; Fig. 5.9, WET) and dissolved oxygen (Fig. 5.10, DRY; Fig. 5.11, WET) from the data presented earlier as profiles. These figures show a few things that were not readily apparent from the profiles presented earlier, so that the present account also contains some new observations.

Although the two groups begin the year with similar profiles of both temperature and dissolved oxygen, the effect of advected heat is soon apparent, especially as the WET years have their greatest mean monthly inflow totals in the first half of the year. Metalimnetic inflows (February - May/June) combine with the outflow centred at c. 45 m to promote downward movement of heat (relative to the surface), effectively broadening the metalimnion and drawing the thermocline down (from a nominal depth of 15 m; ie between 12 m and 18 m) to lie adjacent to the HEPS offtake by May. In DRY years, this part of the annual cycle is characterised by a stable or very slowly deepening thermocline, which moves from a nominal depth of 15 m to a nominal depth of 21 m in May. These events can be seen from a direct comparison of Figs 5.8 and 5.9, remembering that the vertical spacing of the lines is equivalent to gradient in the upper 36 m of the water column where

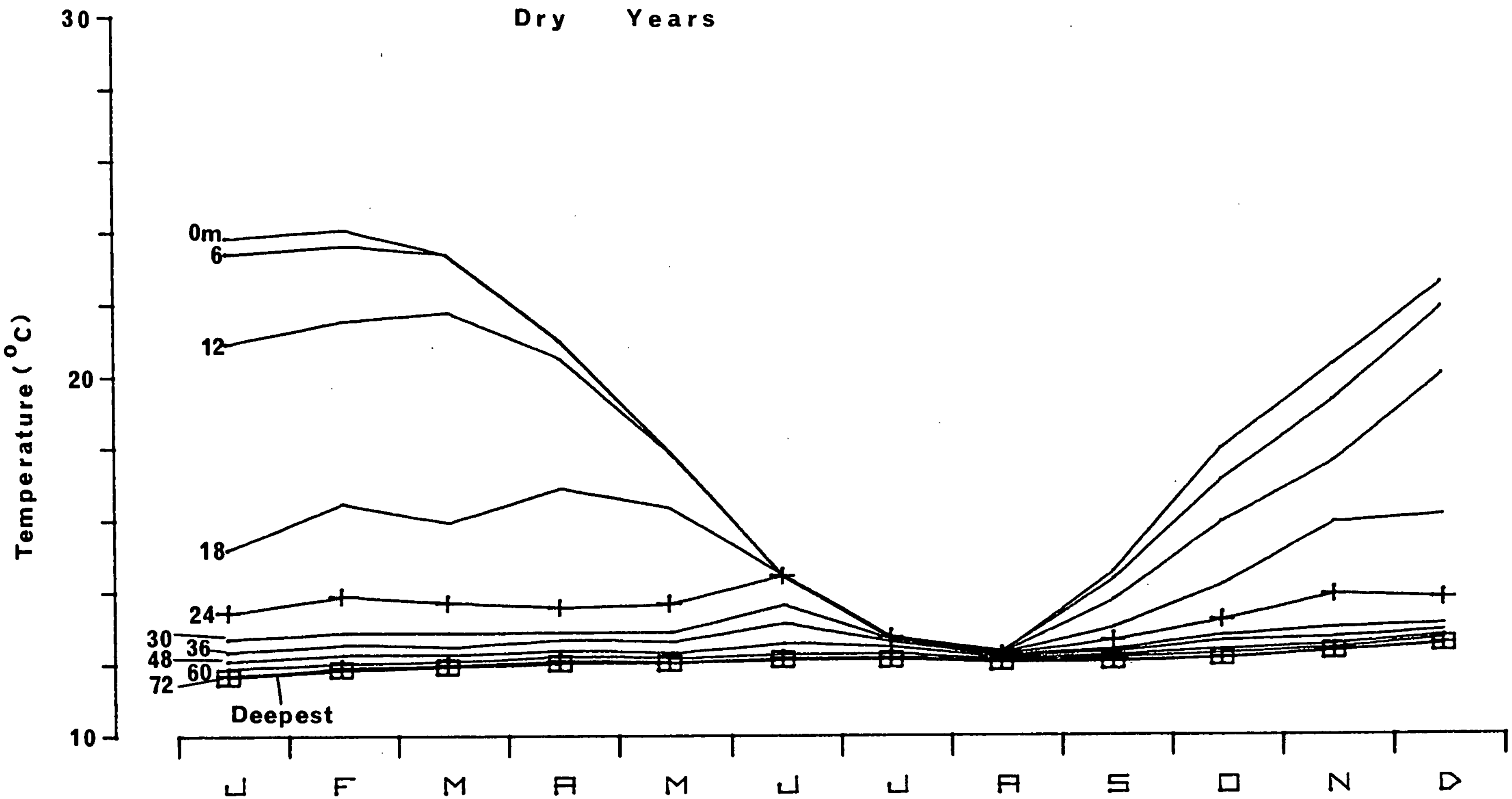
FIGURE 5.8

Seasonal progression of DRY year monthly mean temperatures (°C) for 11 standard depths at site 3D (see Materials and Methods, Chapter 2). This figure essentially combines the separate aspects of profiles and difference plots (see Figs 5.4 and 5.5).

Symbols:-

24 m and 72 m (+)

Deepest sample (□)



Dry Years
Temperature (°C)
30
20
10
0m
6
12
18
24
30
36
48
60
72
Deepest
J
F
M
A
M
J
J
A
S
O
N
D

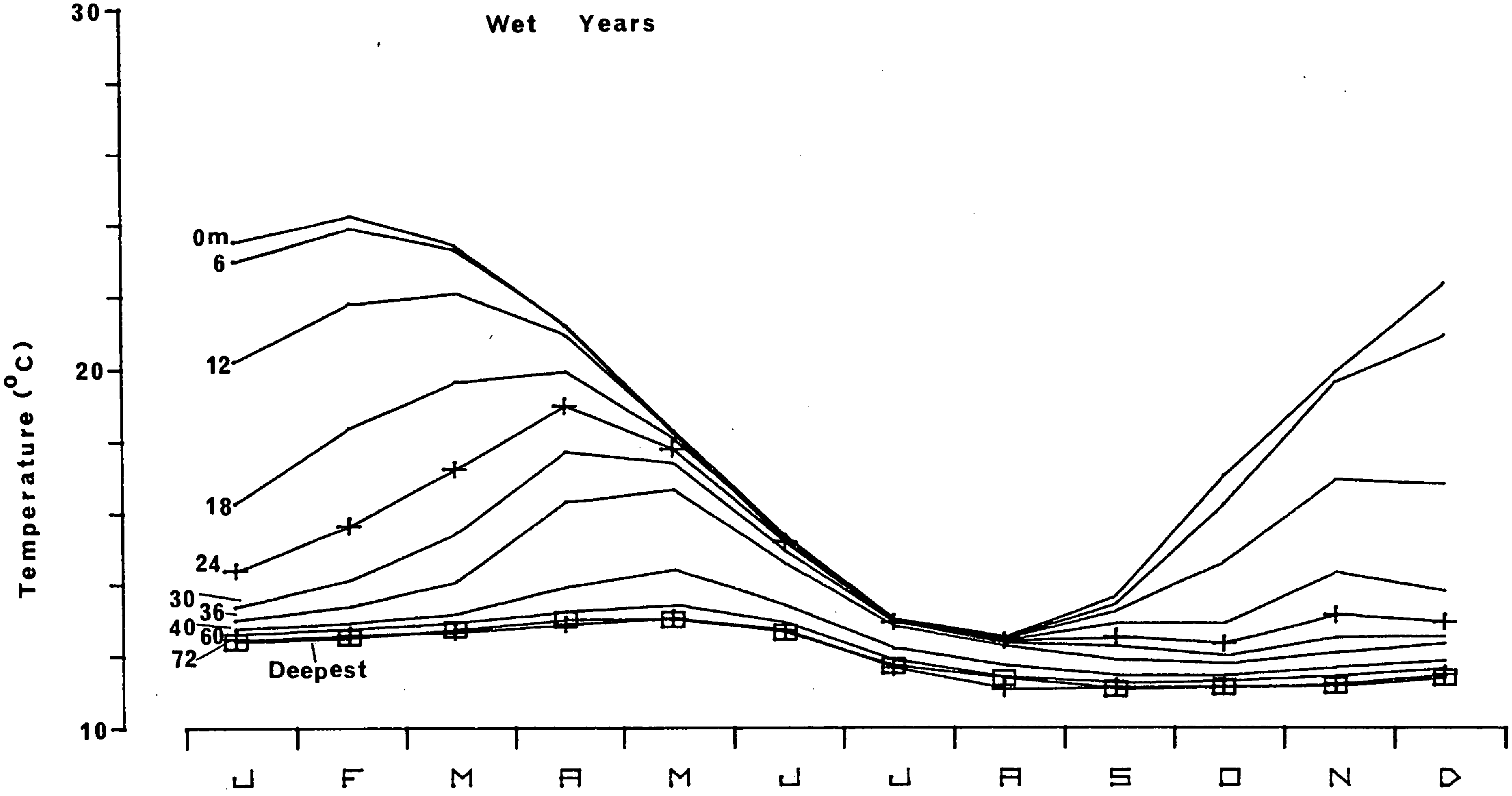
FIGURE 5.9

Seasonal progression of WET year monthly mean temperatures (°C) for 11 standard depths at site 3D (see Materials and Methods, Chapter 2). This figure essentially combines the separate aspects of profiles and difference plots (see Figs 5.4 and 5.5). Compared with the DRY year plots (Fig. 5.8), the greater downward penetration of heat during the first 4 - 5 months, and the cooling effect of winter underflows are particularly obvious.

Symbols:-

24 m and 72 m (+)

Deepest sample (□)



Wet Years
30
20
10
Temperature (°C)
0m.
6
12
18
24
30
36
40
60
72
Deepest
J
F
M
A
M
J
J
A
S
O
N
D

samples are evenly spaced at 6 m intervals. Whole lake heat content and stability (estimated using profiles from site 3D) reflect this enhanced downward heat propagation. Heat content is increased in WET years and the maximum WET year heat content, which occurs later (March instead of February), is 7% greater than the DRY year maximum (% of DRY maximum). In WET years, cooling is delayed by advective heat gain in the first few months of the year, but subsequently proceeds more rapidly than for DRY years (about 50% faster than the 132 cal cm^{-2} day^{-1} DRY year rate, April - July) and the two groups have almost identical winter minimum heat content.

The timing and magnitude of maximum stability are little affected by advection, but the WET year stability is generally less than for DRY years, in this period (January - June). WET year stability is, at most, 500 gm-cm cm^{-2} (33%) less than that of DRY years, in May, and this is also the maximum absolute difference for the year. Birgean wind work is profoundly influenced by advection (which lies outside of its conceptual framework), with WET years showing a delayed (April rather than February), and much greater annual maximum (twice the DRY year maximum).

A comparison of Figs 5.8 and 5.9, indicates that the effect of advected heat is felt at the deepest sampled level at site 3D. The DRY years show an average increase of c. 0.10 °C $month^{-1}$ from January to May, compared to c. 0.15 °C $month^{-1}$ in WET years, despite the fact that both inflow and outflow have their direct effect much higher in the water column at this time of year.

The oxygen stratification, in the first six months, also differs markedly between the two groups. WET years show a volumetric hypolimnetic oxygen depletion rate (VHDR), more than twice the average rate for DRY years. WET year VHDR is 0.92 mg l^{-1} $month^{-1}$ (not temperature corrected, and would be slightly lowered in comparison to the DRY year rate), while the DRY years show a VHDR of 0.44 mg l^{-1} $month^{-1}$. The average period of hypolimnetic oxygen depletion is shortened in WET years, by the direct re-oxygenating

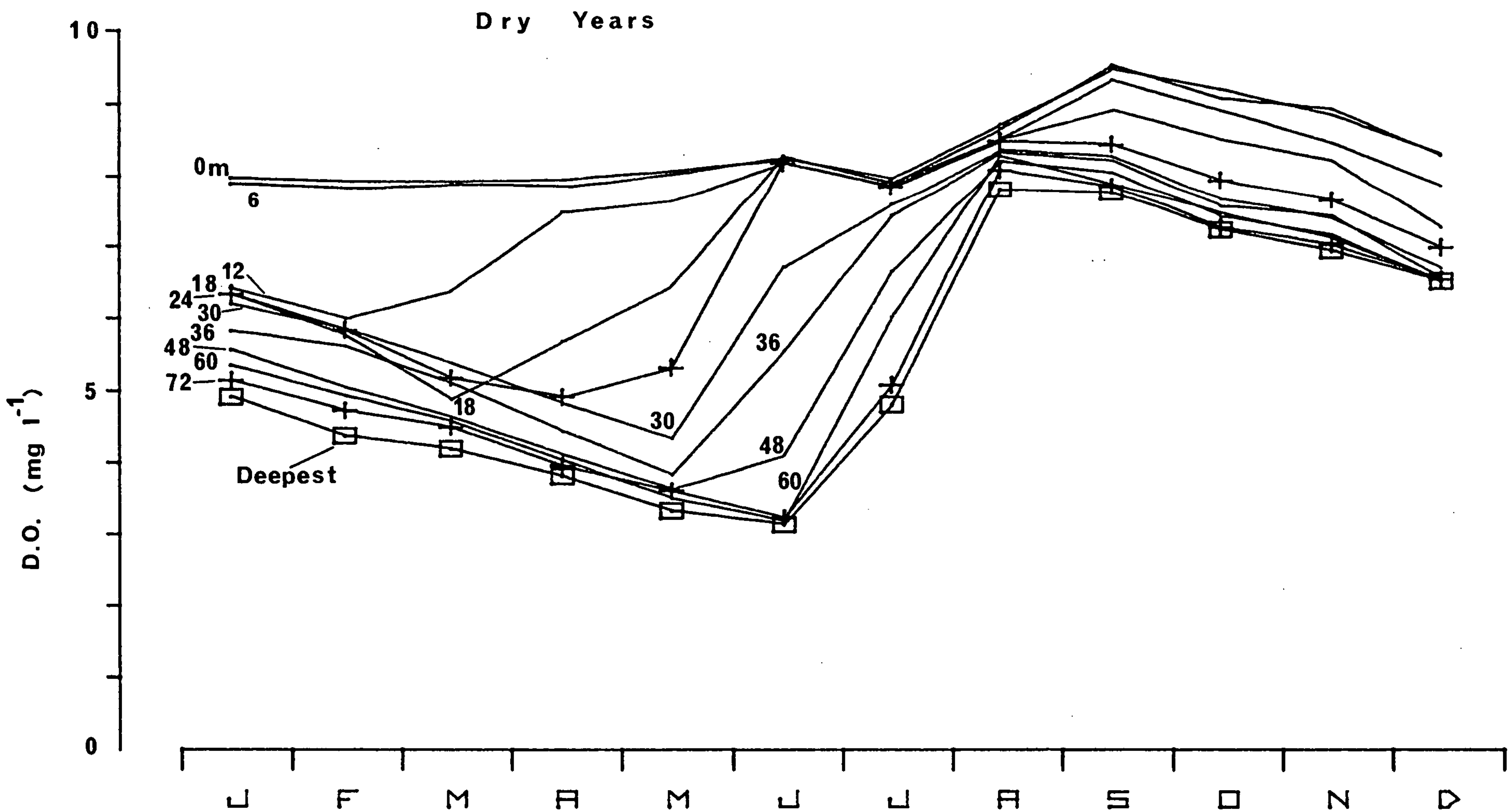
FIGURE 5.10

Seasonal progression of DRY year monthly mean dissolved oxygen (mg l^{-1}) for 11 standard depths (see Materials and Methods, Chapter 2). This figure essentially combines the separate aspects of profiles and difference plots (see Figs 5.6 and 5.7). The formation of a metalimnetic oxygen trough, during the first 3 - 4 months of the year, is evident from the crossing of some lines (particularly 18 m).

Symbols:-

24 m and 72 m (+)

Deepest sample (□)



Dry Years
D.O. (mg l-1)
10
5
0
0m
6
12
18
24
30
36
48
60
72
Deepest
18
30
36
48
60
J
F
M
A
M
J
J
A
S
O
N
D

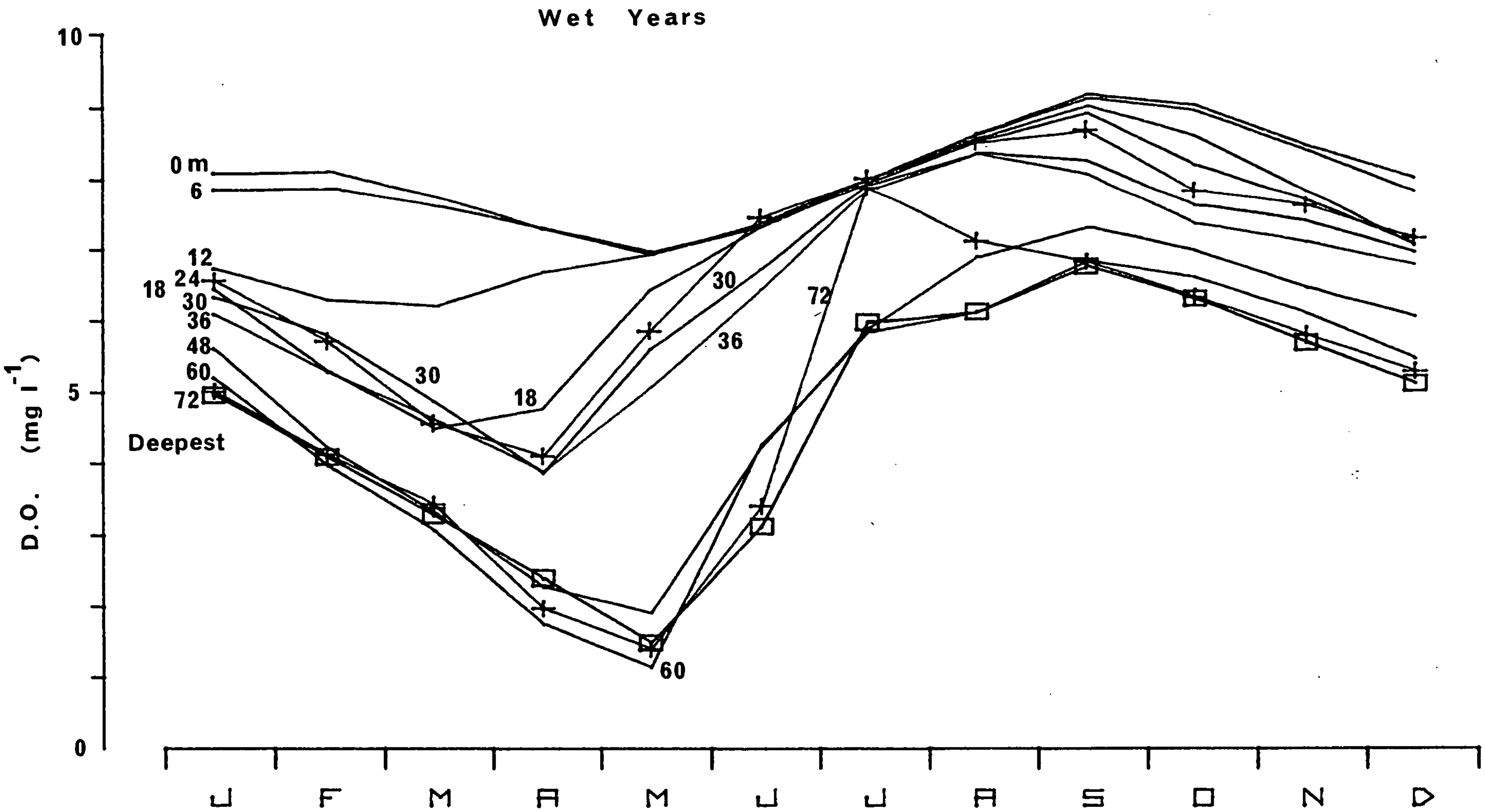
FIGURE 5.11

Seasonal progression of WET year monthly mean dissolved oxygen (mg l^{-1}) for 11 standard depths (see Materials and Methods, Chapter 2). This figure essentially combines the separate aspects of profiles and difference plots (see Figs 5.6 and 5.7). The formation of a metalimnetic oxygen trough, during the first 3 - 4 months of the year, is evident from the crossing of some lines (particularly 18 m). The most obvious differences between the WET and DRY year (see Fig. 5.10) dissolved oxygen data are the greater rate of volumetric, hypolimnetic oxygen depletion, the depression of surface oxygen concentration between February and May of WET years, and the discontinuity that occurs in the profile at the depth of the HEPS offtake (centred at 44.5 m).

Symbols:-

24 m and 72 m (+)

Deepest sample (□)



Wet Years
D.O. (mg l-1)
10
5
0
0 m
6
12
24
18
30
36
48
60
72
Deepest
30
18
30
36
72
60
J
F
M
A
M
J
J
A
S
O
N
D

effect of cold underflows in June. Both groups display a negative heterograde profile in the first three months of the year, though the metalimnetic trough is more pronounced in the WET years. Another obvious feature in Fig. 5.11, is the oxycline that develops adjacent to the HEPS offtake, showing as a broad clear space between 36 m and 48 m (even taking account of the change from a 6 m to a 12 m vertical sampling interval). This clearly contrasts with the DRY years (Fig. 5.10), in which the HEPS offtake is virtually unused.

In the epilimnion, the WET year oxygen concentration declines from February to May, while a slight increase occurs in DRY years. This reflects the generally greater oxygen demand of the more turbid WET year water column. Interestingly, the depth of the convectively well mixed epilimnion, seems almost unchanged in WET years (within the vertical accuracy of this data; ie 6 m), despite the deeper penetration of heat, and the lower stability of the WET year thermal stratification, for instance in May when the WET year water column is 33% less stable than is found in DRY years. This explains the slightly slower rate of surface heat loss in WET years (April - June), because the epilimnetic temperature is a function of both surface cooling, and the incorporation of colder underlying water as the mixed layer deepens. In WET years the deepening epilimnion encounters relatively warmer water than is the case in DRY years, and so remains warmer for a time.

In the period from June to September, the effect of cold underflow is seen in the downward sweep of temperature below 48 m in WET years (Fig. 5.9), compared to a very slight cooling trend in July and August for the DRY years (deepest sample; Fig. 5.8). The minimum thermal gradient occurs in August of both groups and the effect of advection is to maintain a total gradient of 1.2 °C in WET years (none of the WET years, 1963, 1974, 1976, 1978 showed a full vertical circulation), which has a slight discontinuity adjacent to the HEPS offtake. This compare with a monotonic gradient of 0.3°C in DRY years (complete vertical mixing occurred in 1965, 1968, 1980, and probably in 1979 also). This difference extends to the stability of the density distribution which is 3 times greater in WET years (150 gm-cm cm^{-2}), but in neither case is zero stability achieved, on the basis of these averaged monthly profiles. The minimum heat content occurs in August, and is almost the same for both groups. Oxygen concentration in July and August (Figs 5.10 and 5.11) shows a greater gradient (c. 2.5 mg l^{-1}) in WET years, compared with the much smaller DRY year gradient (c. 0.9 mg l^{-1}), and in WET years there is an oxycline adjacent to the HEPS offtake. Unfortunately, however, the WET year data is particularly variable, and incomplete for these months. Epilimnetic oxygen concentration increases in WET years from a minimum in May to its annual maximum in September (Fig. 5.11), whereas in DRY years the epilimnetic concentration, having increased slightly in the first six months of the year, falls between June and July and then increases to the September maximum.

Heating of the surface layers begins in September, leading to the re-formation of a near surface thermocline, and although heat gain is evident in the upper 12 m to 18 m of the water column, both groups tend to lack an isothermal epilimnion (or have an epilimnion < 6 m deep) in the period from September to December. The contrast between WET and DRY years, with respect to inflow and outflow volumes, is less dramatic in this latter part of the year so that the direct effects of advection may be less obvious than in the first part of the year. Nevertheless, some differences are apparent. The WET year hypolimnion, having been cooled by underflows (until September), remains significantly colder than the DRY year hypolimnion, though this makes little difference to the lake heat content because of the small volume involved. Between August and September, surface heating is significantly less in WET years, and although the WET years subsequently catch up this temperature difference, at the surface, heat content is c. 4% - 8% less than that of DRY years in the remaining months. This is mainly caused by the fact that in WET years heating is more superficial than in DRY years, and in December, the WET year thermocline is nominally at 9 m (between 6 m and 12 m), compared to a nominal depth of 15 m for the DRY years. This difference is shown by the vertical spacing of lines (thermal gradient) in Figs 5.8 and 5.9, and may result from the increased turbidity typical of WET years.

Heating rates are approximately linear between September and December, with the DRY year rate (121 cal cm^{-2} day^{-1}), about 10% greater than that of the WET year group. Stability differs only slightly until December, when the WET year group is about 11% less stable. Birgean wind work clearly shows the effect of the delayed heating in WET years, remaining markedly less than for DRY years in this period.

The main features of oxygen stratification between September and December are shown in Figs 5.10 and 5.11. The WET year VHDR is greater than in DRY years, despite the colder hypolimnetic temperatures which continue to decrease until September in WET years. The second oxycline, adjacent to the HEPS offtake, tends to develop in this period (especially November and December), showing some effect of the HEPS withdrawal current, regardless of its less consistent use in the latter part of the year. At the same time, the WET year metalimnion develops a trough (crossing of the 12 m and 24 m lines in Fig. 5.11), and a distinct upper oxycline, between 6 m and 12 m (December, Fig. 5.11), while the DRY year oxygen profile retains much the same form as the thermal profile, with virtually iso-oxic layers near the surface and in the hypolimnion, separated by a single oxycline between 6 m and 36 m (Fig. 5.10).

The Importance of Outflow

In an analysis of the WET and DRY year thermal profiles (January - May), it is possible to compare the extent of downward movement of heat, with the volumes of inflow and outflow. To do this, the the vertical distance between the lower metalimnetic boundaries (arbitrarily, the depth at which the temperature = 14°C) of the WET and DRY year temperature profiles, is assumed to estimate the downward displacement attributable to inflow and/or outflow. The volume between these limits is compared with the volume of total inflow (minus evaporation), and the volume subtracted through the HEPS offtake, in Fig. 5.12. Cummulative totals are plotted because the advective effects on temperature are cummulative. Although the volumes do not coincide, the general shapes of the three curves indicate an almost linear relationship between the volume encompassed by the 14°C positions for WET and DRY profiles and the volume of water subtracted through the HEPS offtake, whereas the cummulative total inflow (minus evaporation) diverges from the other plots after February. Interestingly, the volume between 14°C depths levels off in April, and shows no further increase in May despite continued increases in both of the other volumes (Fig. 5.12). This finding is also consistent with a direct relationship between outflow volume and downward heat propagation, in that the 14°C depth for the WET year profile reaches the centre of the offtake (44.5 m) by April, so that no further downward movement would be expected.

This data indicates that the downward propagation of heat, relative to the lake surface, is more closely related to outflow volume than to the volume of inflowing water, which supports the conclusion that this process is dominated by the vertical bulk movement of water required to balance the loss of water through the offtake, rather than by downward directed turbulence resulting from passage of an inflow through the lake. Further, it was mentioned earlier that the only significant difference in surface temperature, which develops between WET and DRY years in this period (compare June in Figs 5.8 and 5.9), arises because the deepening mixed layer encounters relatively warmer water in the WET years (April - June approximately), and this slows the rate of surface cooling in WET years compared to that of the DRY years. Therefore, the obvious changes in thermal

FIGURE 5.12

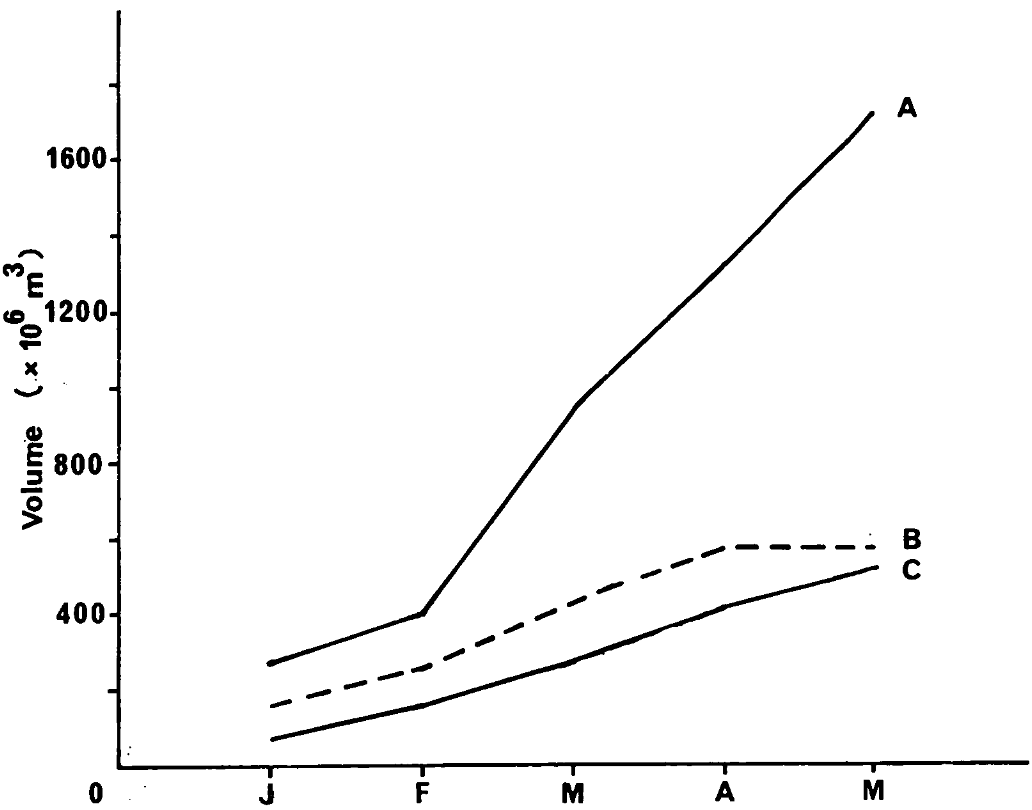
This figure shows the cummulative volumes of total inflow (minus evaporation; 10^6 m^3) and outflow via the HEPS offtake (10^6 m^3) for WET years (the data is taken from Fig. 5.1). These volumes may be compared to the estimated extent of the downward movement of heat for these first 5 months of the year. This has been calculated, as a volume, according to the vertical separation of the WET and DRY year lower metalimnetic boundaries (arbitrarily, the depth at which 14°C is found in each profile; see Fig. 5.4). The cummulative volume, equivalent to this zone of difference between WET and DRY years, is plotted in the figure as a broken line.

Symbols:-

A The cummulative volume of total inflow minus evaporation.

C ~~B~~ The cummulative volume subtracted through the HEPS offtake.

B ~~C~~ The volume equivalent to the vertical separation of WET and DRY year lower metalimnetic boundaries (indicated by the depth at which 14°C occurs in the profiles; see Fig. 5.4).



Volume (×10^6 m^3)
1600
1200
800
400
0
J
F
M
A
M
A
B
C

behaviour (enhanced downward heat propagation and slowed surface cooling), caused by advective processes in WET years, can be accounted for largely in terms of outflow volume rather than inflow volume, within the range of data considered. In other words, the excess of inflow, above that required to balance the subsurface outflow, seems to have relatively little effect on the vertical distribution of temperature at site 3D in this first four to five months of the year (prior to the occurrence of underflows). These comments apply to the excess of inflow, because outflow in the absence of balancing inflow would probably have little effect on the thermal structure relative to the lake surface (see Theory, p 100).

The importance of the outflow may not be surprising, given the proximity of site 3D to the outlet, and it may be reasonable to expect a gradient within a reservoir such as Lake Burragorang, with the thermal structure dominated by inflow induced mixing near to the point of inflow, and by outflow induced downward heat transfer near to the outlet. Coche (1968) draws a clear distinction between the basins of Lake Kariba, depending on their proximity to the inflowing Zambesi River, or to the point of outflow. Bond et al (1978), comment that the operation of offtake structures will dominate the thermal structure in the narrow outlet gorge of Cabora Bassa. However, the focus of the present account is on the site closest to the outflow, and therefore the site of most immediate relevance to water quality management. At this site, the potential effect of advection on the vertical temperature distribution may be approximately predicted from the capacity of the subsurface outflow structure. This applies to the cooling phase when the greatest difference in thermal structure develops between WET and DRY years.

Retention Time and Advective Effects:-

Several points arise out of the preceeding discussion. Firstly, the effects of advection on the temperature profiles (at site 3D) in Lake Burragorang are not simply related to the overall retention time for any year (lake volume/

inflow minus evaporation). A closer approximation to the likely extent of this effect, would come from the ratio of lake volume to the capacity of the subsurface offtake. The HEPS has the capacity to draw off the full lake volume in about 1 year. Consequently, the maximum effect on the thermal structure at site 3D could result from an inflow balanced to the maximum outflow capacity of the HEPS. Under these circumstances, the full inflow volume would pass out of the reservoir at mid-depth, and the annual retention time would be about 1 year. However, the period in which this downward advection of heat is possible does not cover the full year. It is evident from the data presented that a significant change in the thermal profile could be generated from a balanced inflow/outflow over a period of 4 - 5 months (ie January/February to May). If there was little further inflow into the lake for that year (as is possible in this lake), then a retention time of 2 - 3 years might, conceivably, be associated with changes in the thermal stratification of the order of those found in this comparison.

Convection and Advection:-

Secondly, the analysis indicates that there is little discernible effect of inflow (above that needed to balance the outflow), on either the surface temperatures or the depth of the mixed layer in the first part of the year. It may be reasonable to assert that this arises from the time and depth scales used in the comparison. That is, the rate at which temperature is exchanged through the lake surface, between the upper mixed layer and the environment, operates on a comparatively short time scale in relation to the monthly time scale of the analysis. Consequently, any change in the vertical thermal structure, that arises from the bulk upward displacement of epilimnetic water due to the excess of inflow into the metalimnion, is obliterated in a period much shorter than a month and does not affect the profiles averaged over the full month. The present analysis, therefore, detects the changes that occur in the quiescent deeper layers, where heat can be stored away from the rapid exchange processes that operate between the mixed layer and the atmosphere. If this is so, then it implies that modelling of the thermal stratification of such lakes, may be achieved by employing different time-steps to model advective effects than is necessary for characterising the convective effects. That is, if the aim is simply to model the vertical distribution of heat in the lake, it may involve little loss of information to employ a much longer time-step for the advective processes than the one employed to model the convective effects.

I mentioned depth scale as well, and the relatively coarse array of depths used here may well hide significant differences between WET and DRY year profiles, in this period. This is especially so when it is considered that the bulk movement (up or down) caused by a given volume of inflow or outflow varies with depth in the reservoir as a result of the shape of the lake basin. In terms of the vertical temperature profile, which takes no account of volume distribution with depth, the effect of insertion or subtraction of a specified volume of water will be much more readily detected in the deeper layers than in the near surface layers. That is, it must be recognised that the sensitivity of the present comparison to advective effects is least near to the surface, and greatest in the deeper layers. This reflects on the validity of employing vertical temperature profiles in verification of thermal models for such lakes, as seems to be the current practice. Verification of any model that claims to account for advective effects should, therefore, be based on a detailed set of near surface temperature measurments, and include the prediction of lake heat content also.

Outflow and Advective Stabilisation:-

Thirdly, having established the importance of outflow, it is relevant to briefly reconsider advective stabilisation. The comparison of WET and DRY years indicates that a discontinuity in both the thermal and oxygen profiles can be associated with the level of the HEPS offtake (see also Appendix 2).

Wunderlich (1971) describes a similar event in Fontana Reservoir (North Carolina), but does not report any interruption of the circulation pattern. However, Johnson and Merritt (1979) suggest that the outflow current protects the underlying layers from convective circulation in Lake Powell (Utah-Arizona). Welsh (1984) definitely establishes the ability of the high level offtake in Dartmouth Reservoir (Victoria) to confine the convective mixing to the upper water column, even when only 3 m of water overlay the offtake depth. The capacity of the HEPS offtake in Lake Burragorang, is about 90% of that reported by Welsh (1984) for the high level offtake in Dartmouth Reservoir.

I believe, therefore, that there is sufficient evidence to suggest that the HEPS withdrawal layer plays a direct role in vertical partitioning of the water column, and is likely to be important in preventing the penetration of convective mixing to the bottom at site 3D, during the winter period of maximum vertical circulation. Consequently, the process of advective stabilisation has two facets; the stabilising density gradient induced by cold underflow, and the protection afforded by the withdrawal current. Although this aspect was not directly considered in Steane and Tylers (1982) account of Lake Gordon, their paper indicates that the outflow is at a similar level to that reached by convective mixing in 1978 and may, therefore, be significant in the advective stabilisation of this lake also.

A final point, which has not been considered so far, is to what extent these findings apply to the rest of Lake Burragorang. Although I have insufficient evidence to give a rigorous answer to this question, it seems likely that the present account may be directly applied to the Warragamba Gorge section of the reservoir, probably to the 14 km section between sites 3D and Bend (Chapter 2, Figs 2.1 and 2.2). This arm of the reservoir is morphometrically distinct, occupying a particularly narrow and steep sided valley. Beyond the Bend, the reservoir opens out more and the effect of the withdrawal current would presumably be dissipated laterally and by bifurcation into the Wollondilly and Cox River valleys. There is, however, sufficient depth of water to extend an horizontal withdrawal layer from the mid-point of the HEPS at Warragamba Dam to the junction of the Nattai and Wollondilly Rivers without intersecting the bottom of the reservoir (after Soil Conservation Service Report 1962). Considering also, the capacity of the HEPS (capable of subtracting the full volume of water below the upper boundary of the offtake in < 2 months) compared to the small volume of water in the hypolimnion, some effect of the HEPS may be possible over much of the reservoir.

PART II: CONLUDING REMARKS

Re-assessment of Chapters 3 and 4

The analysis of WET and DRY years clearly demonstrates the important effects of inflow and outflow on the fundamental cycles of thermal and oxygen stratification in Lake Burragorang. Advection is also demonstrated to be a very significant factor in the year to year variation of these cycles in the lake. Further, it should be noted that the full range of advectively generated behaviour in the lake has not been covered in the analysis, simply because the comparison af WET and DRY years was constrained to those years which were consistently wet or dry, and did not include years in which a mixture of convective and advective behaviour occurred. Also, the analysis precluded some of the years in which the largest inflows have been recorded (ie 1961 and 1964).

Chapter 5 also permits some resolution of the long term averaged behaviour, described in Chapters 3 (Thermal Stratification) and 4 (Oxygen Stratification), into its component parts. This re-assessment in no way invalidates the earlier descriptive accounts, because the convectively dominated years and the advectively dominated years both contribute to the overall behaviour of the lake. The present discussion may help to separate out the aspects of this behaviour that are best understood within the classic limnological framework built upon studies of natural lakes, and the emerging understanding of reservoir limnology which incorporates some features that are yet to be included in this framework.

The following list briefly sets out some aspects of the lakes behaviour, observed in Chapters 3 and 4, that can be attributed to advective processes:-

1. In Chapter 3 (Fig. 3.5), the seasonal progression of Schmidt stability, heat content, and Birgean wind work showed the three parameters

peaking in successive months. In comparison, for the DRY years (Fig. 5.2) the three parameters all peak in February, while the WET years show successive peaks, similar to those found from the 20 year mean profiles. This indicates that the advection is responsible for the succession of peaks observed in Chapter 3.

2. Advection contributes to the variation of the annual heat budget, by raising the maximum WET year heat content (by 7% of the DRY year maximum), while the two groups show almost identical minimum heat content. Assuming circularity of the data, the WET year heat budget is about 22% greater than that of the DRY years.

3. Following from the previous point, advection increases the cooling rate (April - July) for the WET years to 201 cal cm^{-2} day^{-1}, an increase of 50% of the DRY year rate (132 cal cm^{-2} day^{-1}). Consequently, the assymmetry between heating and cooling rates reported in Chapter 3, is largely a result of advection. For DRY (convectively dominated) years, the rate of heating differs from the cooling rate by less than 10% (of the heating rate). This is similar to the findings for other lakes reported in Table 3.2.

4. Although advection clearly enhances the downward movement of heat at site 3D, during the first 5 months of the year, its effect on the heating efficiency is to decrease it. This occurs because of the very marked increase in Birgean wind work that results from the advective processes. Advection increases the ratio (decreased heating efficiency) by c. 67% of the DRY year ratio (0.144 gm-cm cal^{-1}). This very great sensitivity of the Birgean wind work to advection highlights the restricted application of the whole lake Birgean wind work term, resulting from its dependence on an assumed historic initial condition. But at the same time, probably means that the direct work curve provides a quite sensitive assessment of advective effects (cf. Johnson

and Merritt 1979).

5. In both temperature and oxygen difference plots (Figs 3.4 and 4.6) a downward propagating wave of positive change is observed between about February and May. The observation that "... the zone of negative oxygen change (in the metalimnion) overlaps with a zone of positive heat change ..." (Chapter 4, p 86) is attributable to advective processes. This behaviour does not characterise the DRY years, in which the depth of mixing determines the depth to which the heating wave penetrates.

The occurrence of metalimnetic oxygen troughs is found to be a feature of both WET and DRY years, although it is more exaggerated in the WET years.

<u>Terminology of the interrupted monomixis</u>

There is sufficient evidence for the existence of "advective stabilisation", from Australia, the United States, and probably also from South America (cf. Froelich and Arcifa 1984; I consider their interpretation unlikely), to consider what terminology may be appropriate to describe this interrupted monomictic cycle. This discussion begins with the assumption that it is neither necessary, nor desirable, to erect a separate classification from that applied to natural lakes to describe mixing in reservoir systems. I adopt this position because of the difficulty in drawing a meaningful division between lakes and reservoirs in terms of advective processes (at least in the simple terms of water retention time), especially in relation to lakes like Burragorang that surely traverse, from year to year, the gradient between a system almost totally dominated by convective circulation to one in which advective processes are quite significant. A feature that does separate many reservoirs from most, if not all, natural lakes is the existence of significant sub-surface outflow, and this would be a significant diagnostic feature in a comprehensive classification of lake mixing that attempted to incorporate reservoirs.

Of two possible classifications, "oligomictic" and "meromictic", the recent literature has favoured discarding the former term, and incorporating oligomictic lakes, under the umbrella of meromixis (see Walker and Likens 1975; Lewis 1983). If oligomixis is rejected, then the advective stabilisation must come under the term meromixis, as applied to Lake Powell by Johnson and Merritt (1979). There is no difficulty with this in terms of the definition of meromixis (vertically incomplete mixing), but I feel there is a problem with respect to the generally understood usage of the term meromixis. Even the most transient "spring meromixis" (Åberg and Rodhe 1942; reported in Walker and Likens 1975), involves a prolongation of the isolation of the deeper water layers, and is frequently accompanied by anoxia in the unmixed region (cf. Salonen et al 1984). There is, I believe, an important functional co-occurrence of meromixis and the formation, or maintenance, of anoxic conditions in the monimolimnion, with all the attendant biological consequences that this entails. The following extract from the first conference entirely devoted to meromixis (at Fayetteville, N.Y.) is interesting in this regard:- "The meromictic condition involves a number of important consequences: 1) Absence of turnover deprives the deeper waters of atmospheric contact and leads to progressive deoxygenation with depth. 2) Oxygen depletion at depth typically leads to the production of toxic gases such as hydrogen sulphide and methane. ..." (Carter 1967).

Inclusion of advective stabilisation as a form of transient meromixis introduces a form of meromixis in which the monimolimnion comprises the newest water in the lake (generally well oxygenated, at least initially), and one in which the inflow acts to ensure upward displacement and mixing of the existing hypolimnion. It may be argued that this sequence of events inevitably attends the formation of a type II (sensu Walker and Likens 1975) or triptogenic meromictic lake, but advective stabilisation lacks any catastrophic component being an annual or potentially annual event. Further, it is not at all certain that the turbid nature of these underflows is particularly significant in stabilising the water column, rather, the combination of low temperature and/or the potentially protective role of subsurface withdrawal current seems to be most significant in Lake Burragorang, probably Lake Gordon, and Lake Powell (Merritt and Johnson 1979).

My argument, therefore, is not that advective stabilisation fails to meet the general definition of meromixis, but that the inclusion of this variant effectively uncouples the term meromixis from the generally accepted corollary of monimolimnetic isolation and attendant anoxia which is so significant biologically. Gaining insight into the physical, chemical, and biological phenomena that attend the stratification of lakes is, after all, a major reason for classifying their stratification behaviour. For this reason I consider that although advective stabilisation should be recognised in warm monomictic reservoirs, this behaviour is not adequately described as meromixis and this term should not be applied even as an adjunct in the manner suggested by Lewis (1983). Lake Burragorang, therefore, is considered to be essentially a warm monomictic lake subject to advective stabilisation in about 50% of years.

It is notable, that there have been no formal attempts (that I am aware of) to incorporate the quite diverse stratification behaviour of reservoirs in the limnological scheme of thermal classification, and it may well be that a separate terminology is eventually developed to account, for instance, for what might be called "anthropogenic polymixis" in the case of bubbling de-stratification.

PART III

Chapters 6 - 7

INTRODUCTION

In the following two chapters, some of the water quality parameters monitored at site 3D are examined, particularly turbidity and chloride concentration which are profoundly affected by inflow. The relationship between these parameters and inflow volume is investigated using linear regression. Turbidity, in particular, is considered in some detail because of its relevance to water quality management, both of itself and for its effect on light penetration and nutrient supply which, in turn, affect the phytoplankton biomass (indicated by chlorophyll-a concentration). Measurement of chlorophyll and phosphorus concentration began in 1970, so the consideration of these parameters is confined to the latter half of the study period.

CHAPTER 6

INFLOW

AND

WATER QUALITY

INTRODUCTION

Of the three types of inflow, overflow, interflow, and underflow only the latter two appear to directly affect the water column at site 3D. A number of physico-chemical parameters demonstrate the arrival of discrete inflows at site 3D. Turbidity and chloride concentration are perhaps the most useful indicators, although temperature and dissolved oxygen concentration may also be affected. In the absence of inflow, both the chloride concentration, and more especially the turbidity, are relatively conservative properties of the water column and their use in diagnosing the arrival of inflows at the outlet site has already been mentioned (see Materials and Methods, Chapter 2).

It should be remembered that the visual comparator measurement of turbidity, used here, is neither as sensitive nor objective as the automated system that has replaced it. Present day measurements may be sufficiently sensitive to detect some effect of overflow at site 3D. That these flows can traverse a large lake is evident from the work of Johnson and Merritt (1979) on Lake Powell (U.S.A), which is more than 100 km long and will occupy 300 km of the old Colarado river valley when filled. Here, the summer overflow current is sufficiently large to initiate thermal stratification and is largely responsible for determining the thickness of the summer epilimnion (Johnson and Merritt 1979). Carmack et al (1979) also attribute the initiation of thermal stratification in Kamloops Lake (British Columbia; c. 27 km long) to overflow. In Lake Burragorang, the volume of inflow and/or the temperature relationships between inflows and the lake appear to have precluded any overflow reaching site 3D as a discrete and detectable entity. The largest inflow, in November 1961, reached the outlet site as an underflow, although it occurred at the time of year when overflow is at least possible on the basis of the the temperature data presented in Chapter 3 (Fig. 3.8)

TURBIDITY

"Turbidity is the interference with the passage of light through water by suspended and colloidal matter" (Hart 1974). "While turbidity and the content of suspended sediment are interdependent there is not a direct relation between the two because of (a) the vast range of particle sizes concerned, and (b) the range of particle shapes from spherical to plate-like." (Oades 1982). Nevertheless, turbidity remains a standard water quality measure, especially in the water supply industry (Hart 1974). It might be added here that at a site some 50 km (see Chapter 2, Fig. 2.2) inside a huge particle sorter, which is in effect what a lake achieves by slowing the velocity of influent water, the size range of suspended particles will be considerably reduced. Consequently, the relationship between turbidity and the concentration of suspended sediment may be significantly improved by passage through a lake such as Lake Burragorang.

Turbidity is one of the most obvious manifestations of an inflow penetrating to site 3D. Turbidity is also one of the main indicators of poor water quality in Lake Burragorang, where the chemical deterioration of hypolimnetic water following the development of anoxic conditions is not a primary concern. The contribution of suspended particulate matter to deterioration of water quality arises from its aesthetic unsuitability to the public (including taste, odour and staining of clothes), as well as from problems it may cause in water treatment, by interfering with disinfection and exerting a chlorine demand which makes it difficult to maintain the required residual concentration (Hart 1974). Turbid water may also require a higher level of treatment, such as chemical coagulation or filtration (Linsley and Franzini 1972), which increases the cost of treatment.

In Lake Burragorang some of these difficulties are reduced by the fact that the water is piped to a secondary storage (Prospect Reservoir) prior to its distribution to Sydney. However, feeding turbid water to Prospect may cause problems in that storage, such as increasing the rate of sediment accumulation, and adding nutrients.

Descriptive Account

Isopleths of turbidity for the period 1961 - 1979 (after which the method of measuring turbidity was significantly changed) are presented in Fig. 6.1. Direct comparison with the similarly scaled plot of monthly total inflow (minus evaporation; Chapter 3, Fig. 3.6) readily shows the relationship between major inflows and turbidity at site 3D, and the seasonal distribution of interflows and underflows at this site.

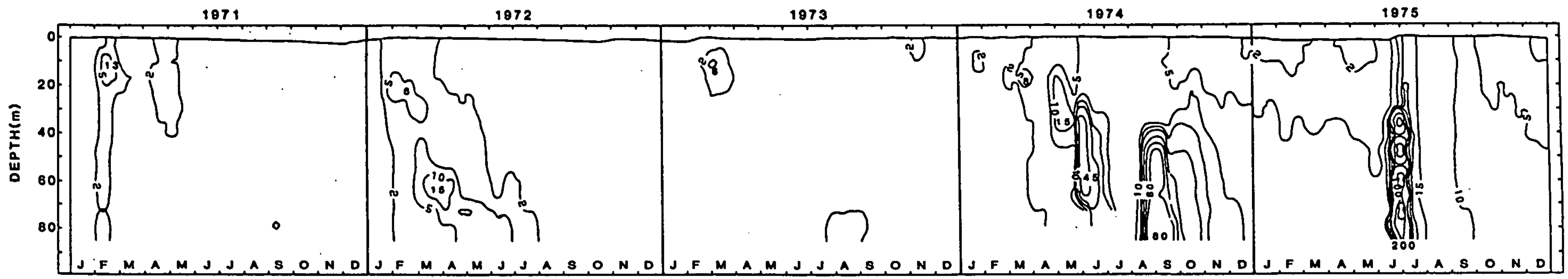
Fig. 6.1 shows that turbidity is generally least near the surface. July and August of 1962 provide an exception to this. During these months a relatively clear underflow displaced the uniformly turbid water left over from the November 1961 flood and resulted in the least turbid water lying at the bottom of the lake.

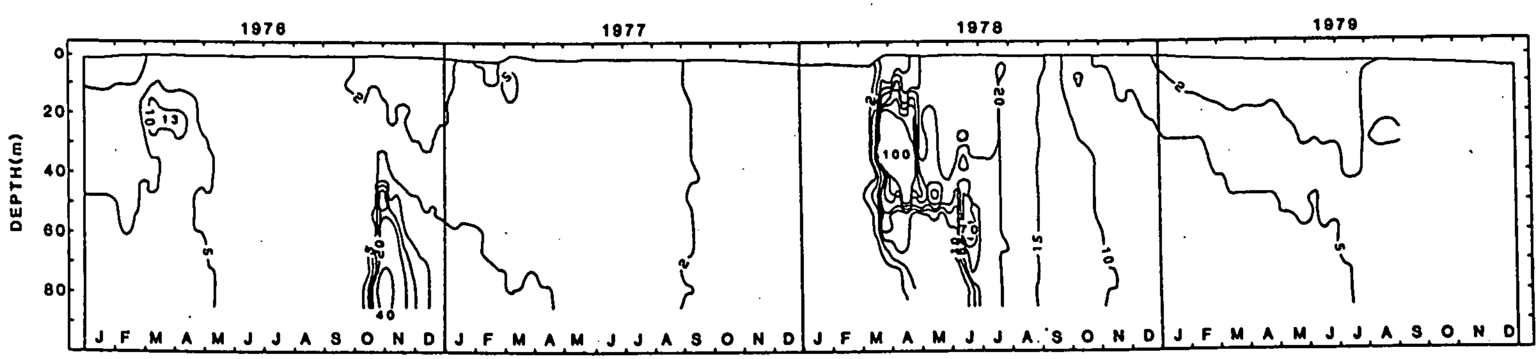
The maximum turbidity recorded for the period 1961 - 1979 was 700 units (Hellige visual comparator), which resulted from the November 1961 flood. The effects of this inflow were still evident 8 months later in 1962, when the July/August period of maximum vertical circulation led to surface turbidity of 50 units. Since 1961, turbidity at site 3D has not exceeded 250 units, which resulted from the August underflow of 1967.

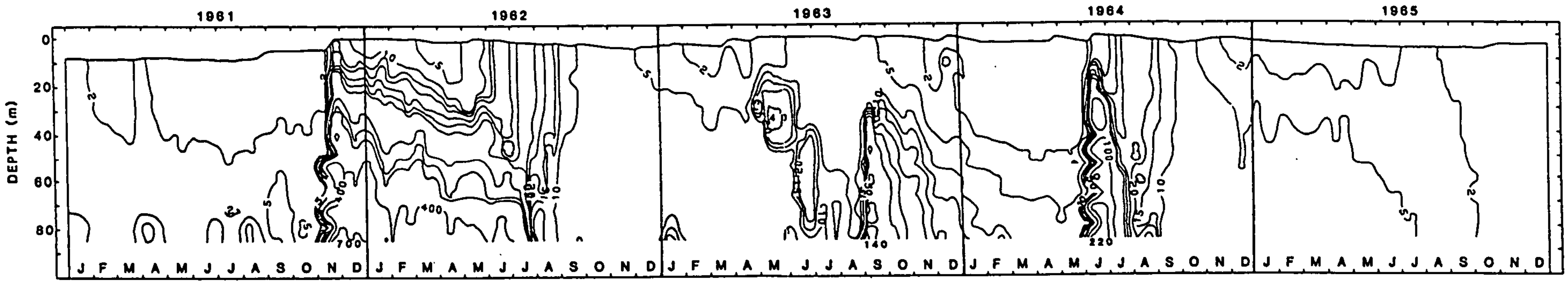
A significant observation from Fig. 6.1, is that interflows have rarely resulted in turbidity greater than 20 units. The maximum record is 100 units for the very large interflow of March 1978. Underflows, on the other hand, may be associated with much greater turbidity at site 3D. This point is shown by Fig. 6.2, in which the maximum turbidity associated with each of the registered inflows is plotted against the total volume of each inflow (see the

FIGURE 6.1

Turbidity (Hellige units) isopleth diagram, 1961 - 1980. Isopleths are drawn at the following intervals:- 2, 5, 10, 15, 20, 30, 40, 50, 100, 150, 200, 400, 600 units. Vertical lines are drawn at the end of each calendar year. Smoothing of sudden changes on the time axis, and the tendency to reduce peak height is inherent in my use of the SURFACE-2 contouring algorithm, and is particularly noticeable for such parameters as turbidity (see Materials and Methods, Chapter 2). To assist in interpreting the figure, the maximum turbidity recorded for each of the major inflows, is marked on the diagram. Most of these turbidity peaks are included in the Inflow Register (Chapter 3, Table 3.3).


1971
1972
1973
1974
1975
DEPTH(m)
0
20
40
60
80
J F M A M J J A S O N D


1976
1977
1978
1979
DEPTH(m)
0
20
40
60
80
J F M A M J J A S O N D


1961
1962
1963
1964
1965
DEPTH (m)
0
20
40
60
80
J F M A M J J A S O N D

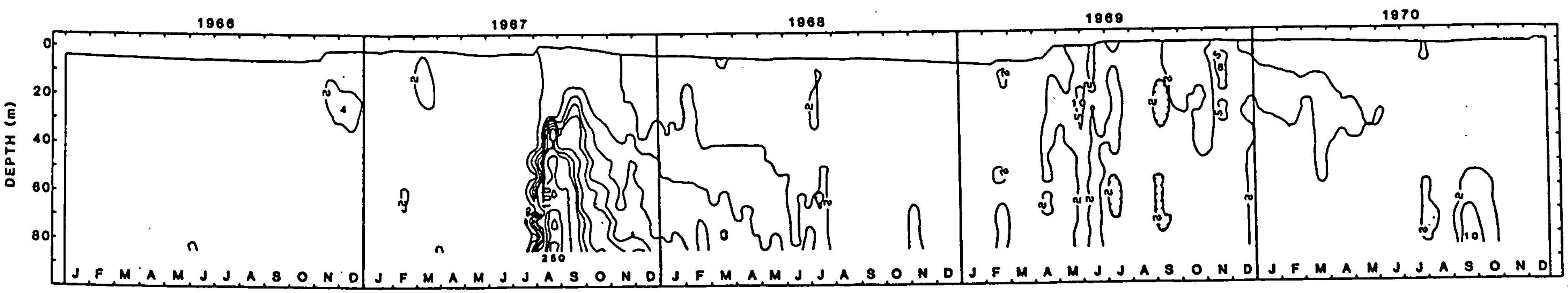

1966
1967
1968
1969
1970
DEPTH (m)
0
20
40
60
80
J F M A M J J A S O N D

FIGURE 6.2

Scatter plot showing the maximum turbidity (Hellige units) associated with registered inflows (see Chapter 3, Table 3.3), that penetrated to site 3D as individually detectable flows in the period from August 1961 to December 1980.

Symbols:-

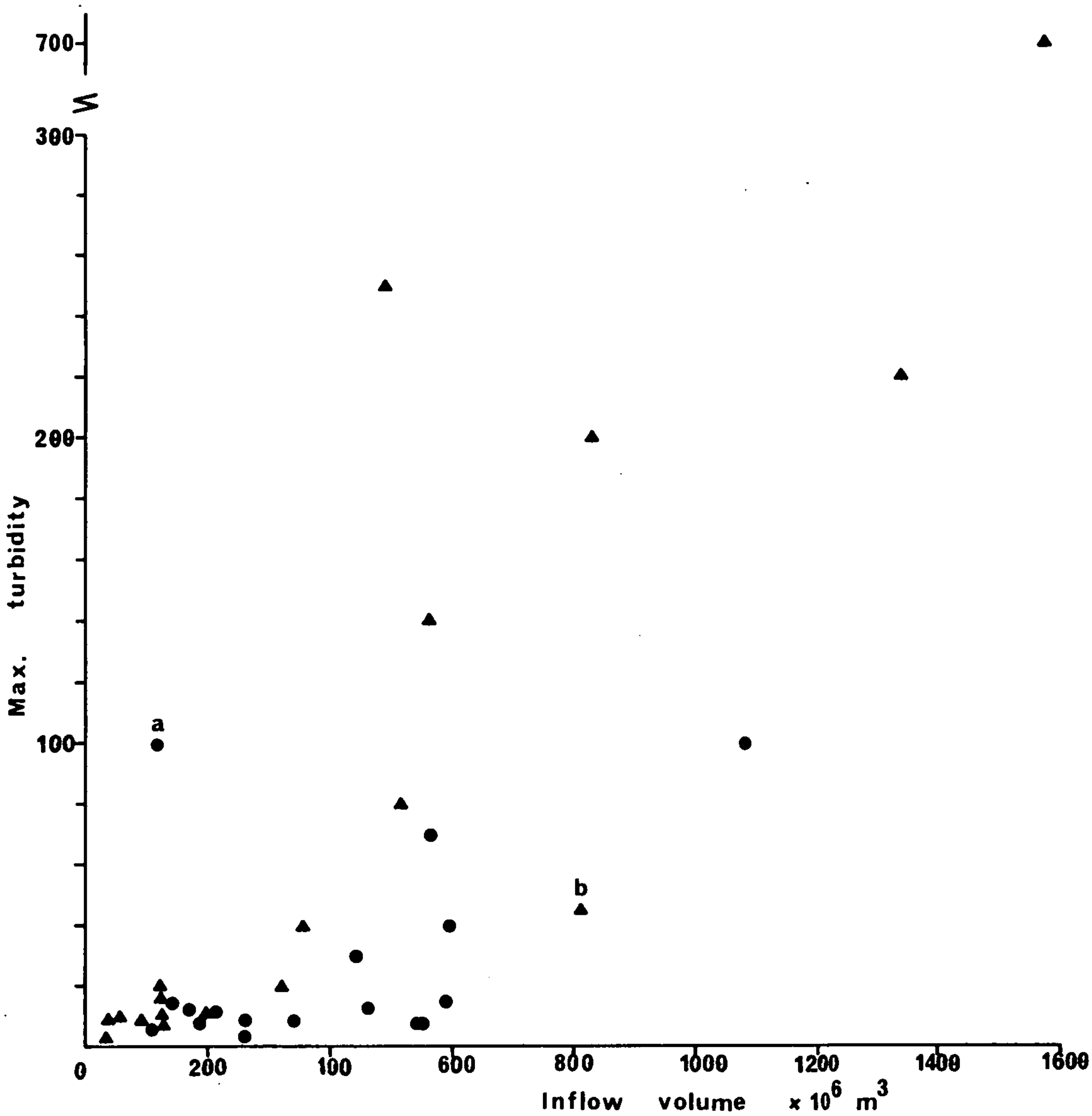
Underflows (▲)

Interflows (●)

Notes:-

a. The April 1978 interflow is uncommonly turbid for such a small volume of inflow. Because this inflow followed soon after the large March interflow, it is possible there was some interaction between the two inflows which resulted in unusually high turbidity per unit volume in the second inflow.

b. The unusually low turbidity of the inflow initiated in May 1974, may be attributed to the fact that the deepest samples taken during this event, were at 60 m below the surface. This would tend to underestimate the maximum turbidity in comparison to other records taken at greater depths.



700
300
200
100
0
Max. turbidity
a
b
200
100
600
800
1000
1200
1400
1600
Inflow volume x 10^6 m^3

Inflow Register; Chapter 3, Table 3.3). It is apparent that underflows are generally more turbid for a given volume than interflows. Two exceptional points are marked on the diagram. The first (initiated in May 1974) has an atypically low turbidity for an underflow. This is probably an artefact of the sampling which was restricted to the upper 60 m of the water column, underestimating the maximum turbidity. Temperature data indicate that the inflow did in fact underflow. The second (initiated in April 1978) may have been affected by the much larger interflow which occurred just before it. Lick (1982), in tank experiments concerning sediment entrainment and deposition, found that recently deposited sediment was easily entrained by turbulent flow. It is conceivable, therefore, that the second and smaller inflow in 1978 picked up some of the recently deposited sediment from the earlier flow. The factors that determine the turbidity of an individual flow are complex and varied, operating within the catchment, the stream, and the lake itself. Consequently, it is best simply to note that the April 1978 interflow seems anomalous in relation to the other data presented here, and not enter into any protracted speculation.

Another observation about interflows, is that they enter the water column over a substantial depth range, from within about 10 m of the surface, to more than 60 m (Fig. 6.1). In years showing successive interflows (ie 1963, 1972, and 1978; also 1974, but the May/June inflow is deceptive, see Fig. 6.1 caption) the later inflow can be seen to affect a lower level; sometimes much lower, for example in 1972.

Underflows tend to produce a greater vertical effect zone, for example in 1964 and 1975 (Fig. 6.1). This can be partly explained in terms of the volume storage curve for Lake Burragorang. More than 90% of the total volume of water in the lake occupies the upper 45 m (below FSL; see Chapter 1, Table 1.2). The extent to which an inflow entrains lake water may also be significant. Imberger and Hebbert (1980) suggest that entrainment of lake water, at the point of entry into the lake, is greater for an underflow than an interflow. Ryan and Harleman (1971; referred to by Slotta (1973) estimate 200% entrainment for an interflow and 500% entrainment for an underflow. A very approximate calculation of the total volume of water markedly affected by turbidity can be made from the isopleths in Fig. 6.1 and the data in Table 1.2. Although this cannot be equated with the entrainment estimates given above, the data indicates that the volume of water affected by underflows is about 90% of the inflow volume, although the figure varies from c. 60 - 120%. This does not take into account the volume of an underflow subtracted through the HEPS offtake and, therefore, represents simply an empirical measure (for 1961 - 80) of the affected volume in Lake Burragorang, assuming the normal operating regime.

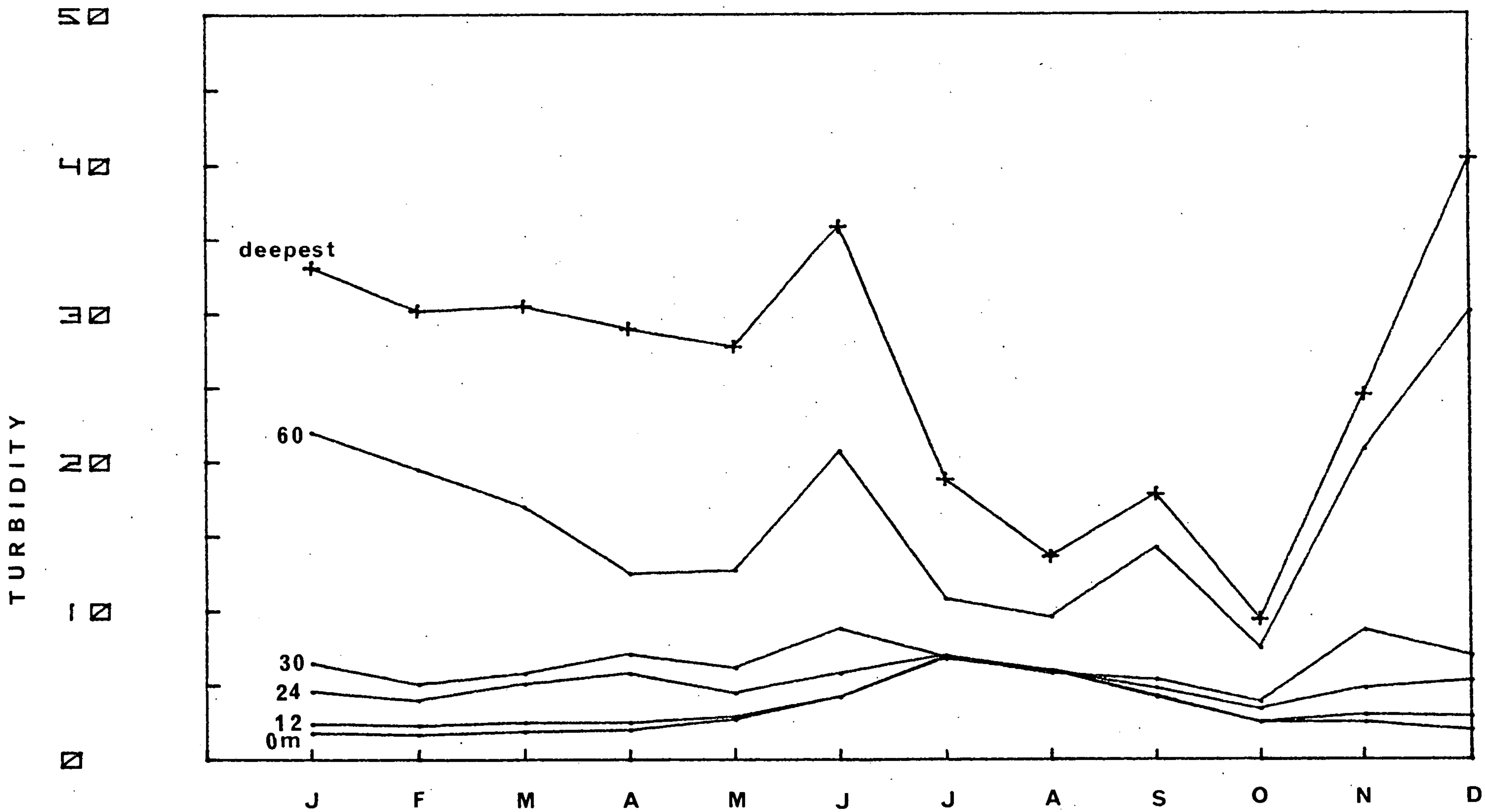
Turbidity in Relation to the Thermal Cycle:-

Major changes in surface turbidity at site 3D are dependent on the timing of inflows in relation to the cycle of thermal stratification and, therefore, the extent of vertical mixing. Maximum surface turbidity is nearly always associated with the July - August period of maximum vertical mixing. Fig. 6.3 shows the monthly mean turbidity derived from data for 1961 - 1979. There is an obvious peak of surface turbidity in the period of maximum vertical circulation (July and August). Stewart and Martin (1982) followed the seasonal pattern of turbidity in Hemlock Lake (N.Y.) and observed that "The entire lake is most turbid during vernal and autumnal circulation and least turbid during winter and summer stratification.". Walmsley (1978) also reports an increased surface turbidity during the circulation period of Lindleyspoort Dam (South Africa).

It is convenient here to regard the cycle of thermal stratification as composed of two phases. The first, in which the extent of vertical mixing is decreasing or comparatively stable and restricted to the superficial water layers, and the second where vertical mixing is increasing and progressively

FIGURE 6.3

Seasonal progression of monthly mean turbidity (1961 - 1979; Hellige units) at various depths. The most important feature of the diagram is the increase in surface turbidity as the mixed depth increases between May and July.


TURBIDITY
50
40
30
20
10
0
deepest
60
30
24
12
0m
J
F
M
A
M
J
J
A
S
O
N
D

recruiting deeper layers. The period September to March approximates the first phase while April to August approximates the second phase. Turbid inflows, either metalimnial or hypolimnial, occurring during or just prior to the second phase (increasing extent of vertical mixing) usually contribute most to surface turbidity. Fig. 6.1 provides examples of this in the March - May interflows of 1963, 1974 and 1978, and the June underflows of 1964 and 1975. June underflows, in particular, have caused very high surface turbidity at site 3D. August, which is one of two months of maximum vertical circulation at site 3D, is an exception. An August inflow might be expected to immediately and drastically increase surface turbidity. However, the occurrence of a turbid underflow is usually accompanied by advective stabilisation of the water column at this time, and this effectively restricts mixing until the onset of thermal stratification in September. The result is a locking up effect such that surface turbidity is only slightly increased. August underflows in 1963, 1967 and 1974 are examples of this. Later underflows (in September, October and November; ie. in October 1976, Fig. 6.1) also have little immediate effect on surface turbidity, unless they are of sufficient magnitude to directly affect the whole water column or they are so turbid that their effect is felt during the next year's period of increased vertical mixing. Both of these comments apply to the November 1961 flood. Interflows during the period of restricted vertical mixing (phase one) are different in that, despite their relatively low turbidity, they commonly cause a small but immediate increase in surface turbidity. This results from the dynamic equilibrium which conditions the vertical extent of the epilimnion, producing limited but frequent water exchange between the upper metalimnion and the fully circulating epilimnion. Fischer and Smith (1983) describe, in some detail, the exchange of water between a plunging inflow and the surface water of Lake Mead, near to the inlet site. They conclude that internal waves, and the formation of temporary thermoclines that are subsequently incorporated into the mixed layer by stronger winds, form important paths of exchange between the inflow and the surface water of the lake. In the case of interflow, these pathways presumably operate further from the inlet site of a lake although at a reduced rate.

Quantitative Relations

Linear regression analysis is used here in an attempt to find predictive relationships between the inflow and the maximum turbidity recorded at site 3D. Also, because the surface water is usually least turbid, prediction of the most turbid condition of the upper water layers is useful in managing the lake. In many instances logarithmic transformation of the variates was necessary to meet the requirements of the linear regression model. Those regressions reported without transformation, met the requirements of the model in arithmetic units (see Materials and Methods, Chapter 2). Unfortunately, data concerning the inflow volume prior to December 1963 became available after the bulk of the analyses were performed and is not included here, except for a few of the more significant relationships.

Maximum turbidity and inflow:-

The data for this analysis comes from the Inflow Register (Chapter 3, Table 3.3). The underflow initiated in June 1978 is excluded from consideration as it produced no change in the ambient turbidity at site 3D. The data is analysed firstly as a whole, and then the subsets of interflows and underflows are examined separately.

The complete data set (15 interflows and 10 underflows; from the interflow of December 1963 to the interflow of June 1978, Table 3.3) indicates that there is an average linear relationship between the maximum turbidity (recorded at 3D) and the inflow volume for the specific event, and that the slope of the regression line is significantly different from zero at the $p = 0.01$ level (F statistic from regression ANOVA). The regression does not provide a good predictive equation, as the independent variate (total inflow) accounts for less than 40% of the variation in the maximum turbidity at 3D (r^2= 0.37; n = 25: $100r^2$= % of variation in Y explained by its regression on X). The deliberate exclusion of the data for April 1978 interflow improves the linear relationship such that 48% of the variation of maximum turbidity is explained by its regression on the total inflow volumes for major inflows (r^2= 0.48; n = 24). This still does not provide the basis for a good predictive equation.

It was suggested earlier that there is a different relationship between turbidity and inflow volume depending on whether the inflow penetrates to site 3D as an interflow or an underflow. Because the type of inflow depends on the form and extent of thermal stratification, it is possible that some numerical measure of thermal structure may account for some of the unexplained variation in the above regressions. A numeric assessment of the state of thermal stratification in a lake is possible using the lake stability concept. Data for the Schmidt stability (gm-cm cm^{-2}), derived from the temperature profile taken nearest to the initiation date of each inflow, has been included in a multiple regression of maximum turbidity on the two independent variates, the specific inflow volume and the Schmidt stability. The inclusion of the stability factor contributes significantly to the earlier regression. The multiple regression explains 66% of the variation of maximum turbidity on specific inflow volume (n = 24; excludes April 1978), an improvement of 18%. The resultant equation is:-

ln Turb =	1.085 ln Vol	- 0.394 ln S	- 0.481	(F = 20.5, p = 0.00001)
SE	0.181	0.116	1.188	

The turbidity is negatively correlated with the Schmidt stability immediately prior to the inflow, and positively correlated with the inflow volume.

Maximum Turbidity and Interflows:-

Partitioning the data into the two component inflow types, and excluding the data for the April 1978 interflow, leaves fourteen interflows for the period from December 1963 to 1978 (see Table 3.3).

Inflow volume explains 33% (r^2= 0.33) of the variation in maximum turbidity associated with interflows. Although the F statistic indicates that the relationship is significant at the 5% ($p = 0.05$) level, predictive power is minimal. It is possible that the maximum interflow turbidity might be related to the lake stability, in that a lower stability could be indicative of an interflow penetrating site 3D at a greater depth. Such an interflow would travel further into the lake before "lifting off", and may be more turbid. However, addition of the stability data added almost nothing to the percentage of variation in turbidity explained by the regression.

Maximum Turbidity and Underflows:-

A highly significant linear relationship exists between the maximum turbidity generated by underflows and the volume of inflowing water, for the 10 underflows in the data set. The regression explains 84% of the variation in the maximum recorded turbidity at 3D, and offers some predictive potential, despite the small number of data pairs. Fig. 6.4 shows the least squares regression line with boundaries for 95% confidence limits and 95% prediction limits (for prediction from single values of the independent variate; inflow volume), the axes are given in ln units. The average relationship is described by the following equation:-

$$\ln Turb_u = 1.127 \ln Vol_u - 2.673 \quad (F = 43.2,\ p = 0.0002)$$

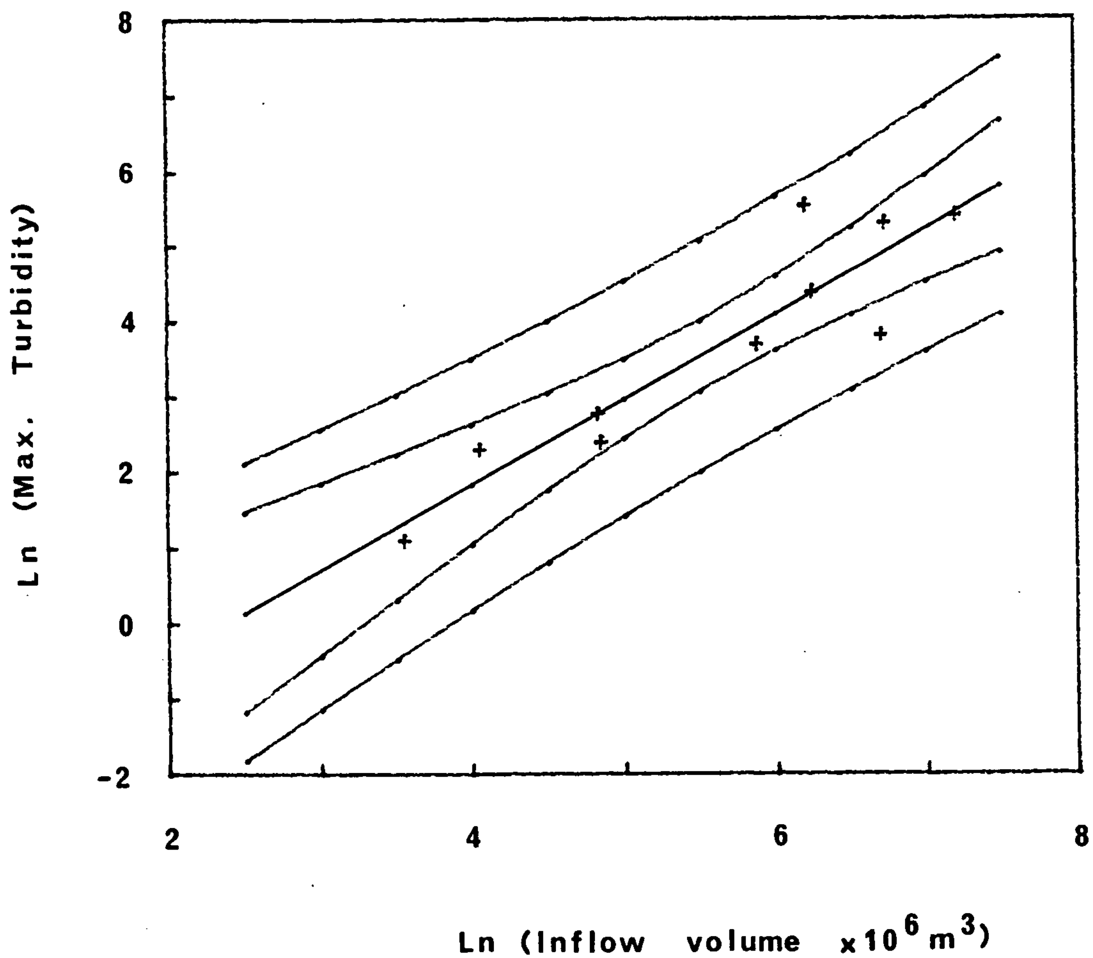
SE	0.171	0.985

Interestingly, the inclusion of data back to August 1961, which adds another 7 underflows (see Table 3.3), has little effect on the regression equation. The slope and intercept remain within 1 SE of the values given

FIGURE 6.4

Linear regression of maximum turbidity (Hellige units), associated with underflows, on the total inflow volume (10^6 m^3), for data pairs from the inflow register (Chapter 3, Table 3.3). Both variates are ln transformed. Boundaries for 95% confidence (innermost lines) and prediction (outer lines) limits (for prediction from single values of the independent variate; Inflow volume) are marked on the plot. The regression explains 84% (100 r^2) of the variation in the maximum recorded turbidity at site 3D. (n = 10 underflows).

$\ln Turb_u$ =	1.127 $\ln Vol_u$	- 2.673	(F = 43.2, p = 0.0002)
SE	0.171	0.985	



8
6
4
2
0
-2
2
4
6
8
Ln (Max. Turbidity)
Ln (Inflow volume x10^6 m^3)

above. It was mentioned earlier that the underflow initiated in May 1974 could have underestimated the maximum turbidity because the greatest sampled depth was 60 m. Exclusion of this point, leaving 16 data pairs, also has only a small effect on the regression equation, and does not change the percentage of variation explained by the regression.

Discussion:-

A variety of factors probably influence the ultimate turbidity levels recorded at 3D, in response to a given underflow. These will include factors relating to the duration and intensity of rainfall, the area subject to the greatest volume and intensity of rainfall within the 9000 km^2 natural catchment, the state of vegetative cover in relation to season and fire, and the previous rainfall history.

Olive and Walker (1982), in a review of the processes that produce suspended sediment in water, conclude "The response of suspended loads to storm events is obviously so complex that simple models using rainfall, geology, soils and vegetation factors are unlikely to account for the variations that occur.". It is of considerable interest and practical significance, then, to find that such a high percentage of the variation in maximum turbidity can be accounted for simply by the volume of underflows. It is probable that the scale of this relationship is so large as to average out the smaller scale variability, in much the same way Harris (1985) argues is the case for relationships such as that between total phosphorus and chlorophyll-a concentrations in lakes.

Considering that underflows generally contribute most to the turbidity and associated water quality problems at 3D, the prediction of their effect may be of particular importance in managing the lake. This utility would, however, be considerably enhanced if the independent variate (inflow) was related to or replaced by some measure, such as the maximum daily inflow, which may occur early in the inflow event and provide an estimate of the maximum turbidity prior to its occurrence.

A comparison of the interflow and underflow regressions confirms the observation that interflows are associated with lower maximum turbidity for a given inflow volume (arithmetic predictions of maximum turbidity resulting from an inflow of 800 10^6 m^3 are 26 for an interflow and 129 for an underflow). Various reasons may be advanced to explain this. First, an interflow must at some point lift-off and intrude into the metalimnion of the lake. At this point, the gravitational component of the inflow velocity is reduced compared to an underflow, which flows downhill to the dam in a lake like Burragorang. Also, the inflow loses some area of contact with the lake floor (it may retain contact with the lake sides). These effects decrease the velocity of the inflow, leading to increased settling out of the suspended sediment carried by the flow, and substantially reduce the source of sediment, preventing replenishment of the suspended load by entrainment of sediment already deposited on the lake floor (cf. Lick 1982).

In relative terms, then, an underflow into Lake Burragorang probably has a greater velocity, and therefore ability to carry suspended sediment, and is also able to entrain sediment already deposited in the lake as it flows towards the dam. The ability of underflows to scour out channels in the sediment of the lake floor is well known (Pickrill and Irwin 1982). The velocity of an interflow after it enters the metalimnion may be very slow, depending on the longitudinal slope of the isotherms (Neel 1963). Whereas the underflow proceeds downhill, it is possible that the isotherms will be tilted in a direction opposed to the passage of an interflow. Bowmaker (1976) reports that the isotherms tilt up towards the dam in Lake Kariba; this must impede the passage of interflows towards the dam.

Coriolis deflection (to the left in the Southern Hemisphere), may further delay the passage of an interflow to the outlet site, while an underflow is channeled via the old river bed to the dam. Also an interflow enters a layer in which its volume is able to spread out more, laterally, contributing further to its dissipation in the main body of the lake. Finally, an interflow enters and travels in a considerably more dynamic zone than an underflow, and there is presumably greater chance of the interflow being scavanged by temporary downward incursions of the mixed layer (cf. Fischer and Smith 1983).

Surface Turbidity and Inflow

The relationship between maximum surface turbidity and inflow is examined using annual data, because the effect of an inflow on surface water turbidity is closely related to the timing of the inflow in relation to the thermal stratification cycle. These analyses do not use information from the inflow register.

Two independent variates are used to predict the maximum annual record of surface turbidity, first, the annual total of inflow minus evaporation, and second, the maximum monthly inflow divided by the mean inflow for the six preceeding months is used in an attempt to measure of the intensity of the maximum monthly inflow for each year. This annual flow intensity index is designed to distinguish between inflows which follow extended dry periods, and inflows which form part of a series of relatively large inflows. Cooke and Williams (1973) report that the first flow after a dry period is often turbid. Also, the process of sediment exhaustion can result in progressively lower turbidity in a succession of flows (Olive and Walker 1982). Sixteen years of data have been analysed (1964 - 1979)

A multiple regression of maximum surface turbidity against the total annual inflow and the calculated intensity factor rejects the intensity factor as contributing little to the overall regression. The simple regression of maximum surface turbidity on the total annual inflow accounts for 50% of the variation of surface turbidity and the slope of the relationship is significantly different from zero at the 1% ($p = 0.01$) level of probability. The percentage of variation explained by the regression is not high enough to give a good predictive equation. The inflow intensity factor is found to be significantly correlated with annual inflow at the 5% level of probability, and although it accounts for some 30% of variation in surface turbidity it is not truly independent of the annual inflow and consequently fails to fulfill its intended purpose. Addition of the two data pairs for 1963 and 1962 results in a weakening of the regression, so that it explains 44% of the variation in maximum surface turbidity. The regression remains significant at the 1% ($p = 0.01$) level.

The attempt to provide a quantitative analysis of the factors contributing to surface turbidity maxima has two aspects. First, the surface is commonly the least turbid, and therefore best quality water, taking only physico-chemical factors into account. Second there is a demonstrable connection between surface turbidity and factors which influence the growth of phytoplankton in the near surface region of the lake. The effect of turbidity on light extinction is well documented (see Kirk 1977b, 1982), and a highly significant linear correlation can be demonstrated between turbidity and secchi disk transparency in Lake Burragorang. Further, there is a close association between suspended particles (and/or colloids) and algal nutrients, particularly phosphorus (Cullen and Rosich 1979; Oades 1982). Surface turbidity is to some extent correlated with annual mean total phosphorus in Lake Burragorang. Finally there is a significant correlation between the measured concentrations of total phosphorus and Chlorophyll-a in Lake Burragorang. These inter-relations will be more fully discussed shortly.

TOTAL IRON

Hart (1974) describes iron as a "... particularly objectionable constituent in water supplies for either domestic or industrial use.". Iron may affect the taste of water, and can stain clothes and bathroom fixtures (Hart 1974). Incrustations of ferric hydroxide can also form in water mains (Linsley and Franzini 1972). The recommended concentration of total iron in water for domestic supply is 0.3 mg l^{-1} or less, according to Hart (1974). Much greater concentrations have been recorded for Lake Burragorang after floods. The concentration of total iron reached 40 mg l^{-1} following the November 1961 flood. Because of the obvious co-occurrence of inflow and increased concentration of total iron, the relationship between total iron and turbidity is investigated for registered inflows (Chapter 3, Table 3.3). There is a close linear relationship between the maximum turbidity (Table 3.3) and the matched record for total iron concentration, using all the registered inflows for which a change in turbidity was recorded (n= 36; excludes underflow initiated in June 1978). The regression explains 82% of the variation in concentration of total iron, and provides a reasonable predictive equation:-

$$\begin{array}{lll} \ln Fe_{tot} = & 0.920 \ln Turb - & 2.778 \quad (F = 155; p = 3.2\ 10^{-14}) \\ SE & 0.074 & 0.251 \end{array}$$

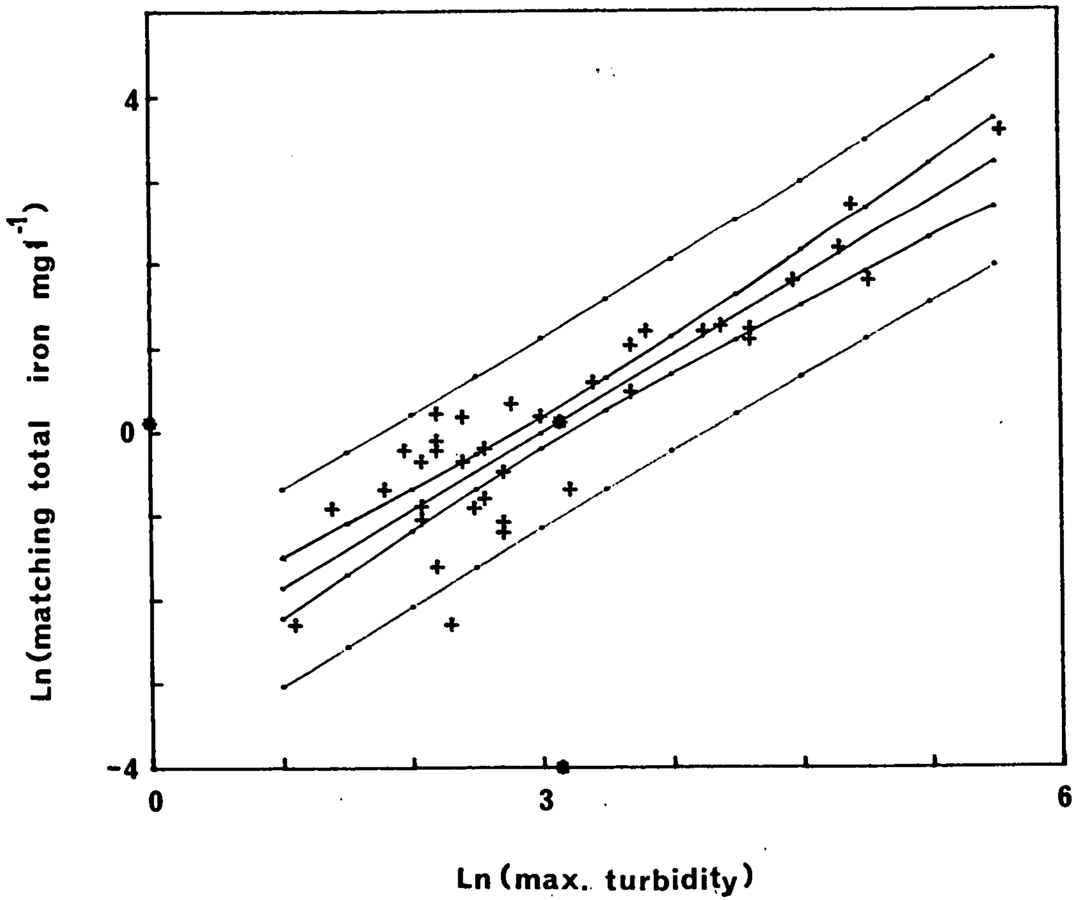
This regression is shown in Fig. 6.5, with 95% confidence and predictive limits. In view of the fact that the Hellige turbidity is no longer determined, this relationship is important in that it offers some potential for transposing the old turbidity measurements into nephelometric units, if a close relationship is found between the total iron concentration and turbidity measured by the new method. Reversing the regression, making total iron the independent variate, gives the following equation:-

$$\begin{array}{lll} \ln Turb = & 0.891 \ln Fe_{tot} + & 3.041 \quad (F = 155; p = 3.2\ 10^{-14}) \\ SE & 0.072 & 0.094 \end{array}$$

FIGURE 6.5

Linear regression of total iron concentration (mg l^{-1}) on maximum turbidity (Hellige units), for turbidity data from the inflow register (Chapter 3, Table 3.3), and samples of total iron taken at the same time. Both variates are ln transformed. Boundaries for 95% confidence (inner) and prediction limits (outer) are marked on the plot. The regression explains 82% of the variation in concentration of total iron (n = 36; excludes the underflow initiated in June 1978).

$\ln Fe_{tot}$ =	0.920 ln Turb	− 2.778	($F = 155$; $p = 3.2\ 10^{-14}$)
SE	0.074	0.251	



4
0
-4
0
3
6
Ln (matching total iron mg l^{-1})
Ln (max. turbidity)

Brymner (1983) reports a close correlation between iron and turbidity in Lake Hume (Victoria). From the above regressions it seems likely that the increase in total iron associated with turbid inflows takes the form of bound, particulate iron and not soluble iron. Pik et al. (1982) found that 90% of iron was in the particulate fraction in the Hawkesbury estuary downstream of Lake Burragorang.

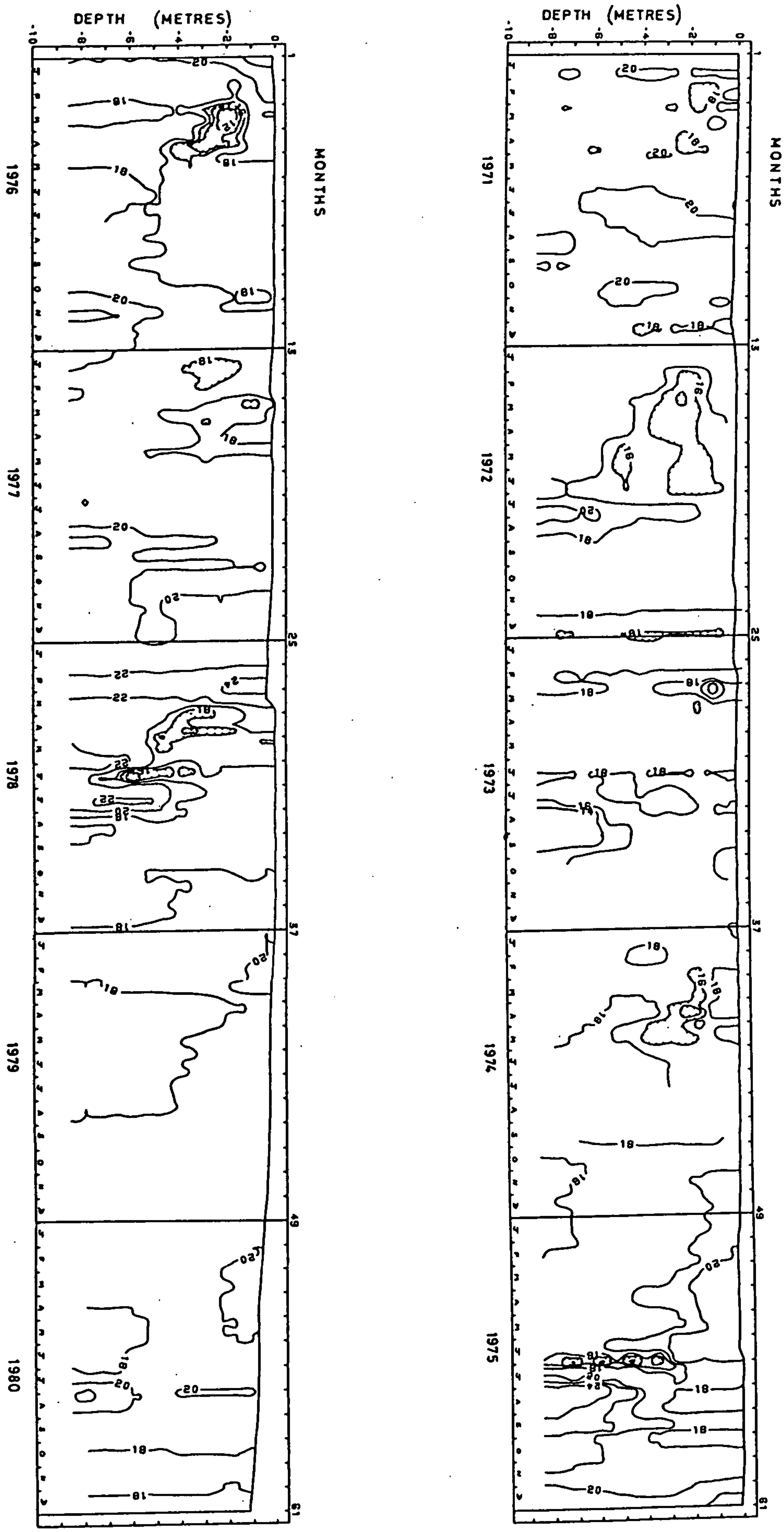
CHLORIDE

Isopleths of chloride concentration (mg l^{-1}) are shown in Fig. 6.6. Comparison of Fig. 6.6. and the monthly total inflow (see Chapter 3, Fig. 3.6) clearly shows that inflows commonly produce changes in chloride concentrations at site 3D. The changes can be in either direction from the normal concentration of about 18 - 20 mg l^{-1}. Two examples are 1961, when the November inflow of dilute water lowered chloride concentrations at 3D to c. 7 mg l^{-1}, and July 1963 when a lesser inflow of more concentrated water caused concentrations to rise to 32 mg l^{-1} in the lower strata (Fig. 6.6).

The ability of an inflow to increase or decrease chloride concentration at site 3D is related to the distinction in water chemistry between the Cox and Wollondilly Rivers. This distinction is probably explained by the different geology of the respective catchments (see Chapter 1). Scatter plots of alkalinity (mg l^{-1} as $CaCO_3$) against chloride (mg l^{-1}; Fig. 6.7) demonstrate the chemical difference between the two rivers. The data comes from the inflow sample sites for the two major rivers that flow into Lake Burragorang (see Chapter 2, Fig. 2.1). The three years shown in Fig. 6.7 represent almost the full range of total annual inflow (minus evaporation) for the study period. These are, in descending order, 1974 (3904.5 10^6 m^3) one of the wettest years for the study, 1973 (881.8 10^6 m^3), and 1979 (98.9 10^6 m^3) which was

FIGURE 6.6

Chloride (mg l^{-1}) isopleth diagram, 1961 - 1980. Isopleths are drawn at the following intervals:- 10, 12, 14, 16, 18, 20, 22, 24, 26, 28, 30, 32, 34 mg l^{-1}. Vertical lines are drawn at the end of each calendar year. Closed depressions of chloride concentration, are hatchured to contrast readily with peaks of chloride concentration. It is evident, that inflows are associated with both increases and decreases in chloride concentration, relative to the ambient concentrations at site 3D. This can be shown, in part, to be a product of the relative proportion of the inflow volume, contributed by each of the two main inflowing streams, as these streams (the Wollondilly and Cox Rivers) are chemically distinct.



DEPTH (METRES)
MONTHS
1976
1977
1978
1979
1980
DEPTH (METRES)
MONTHS
1971
1972
1973
1974
1975

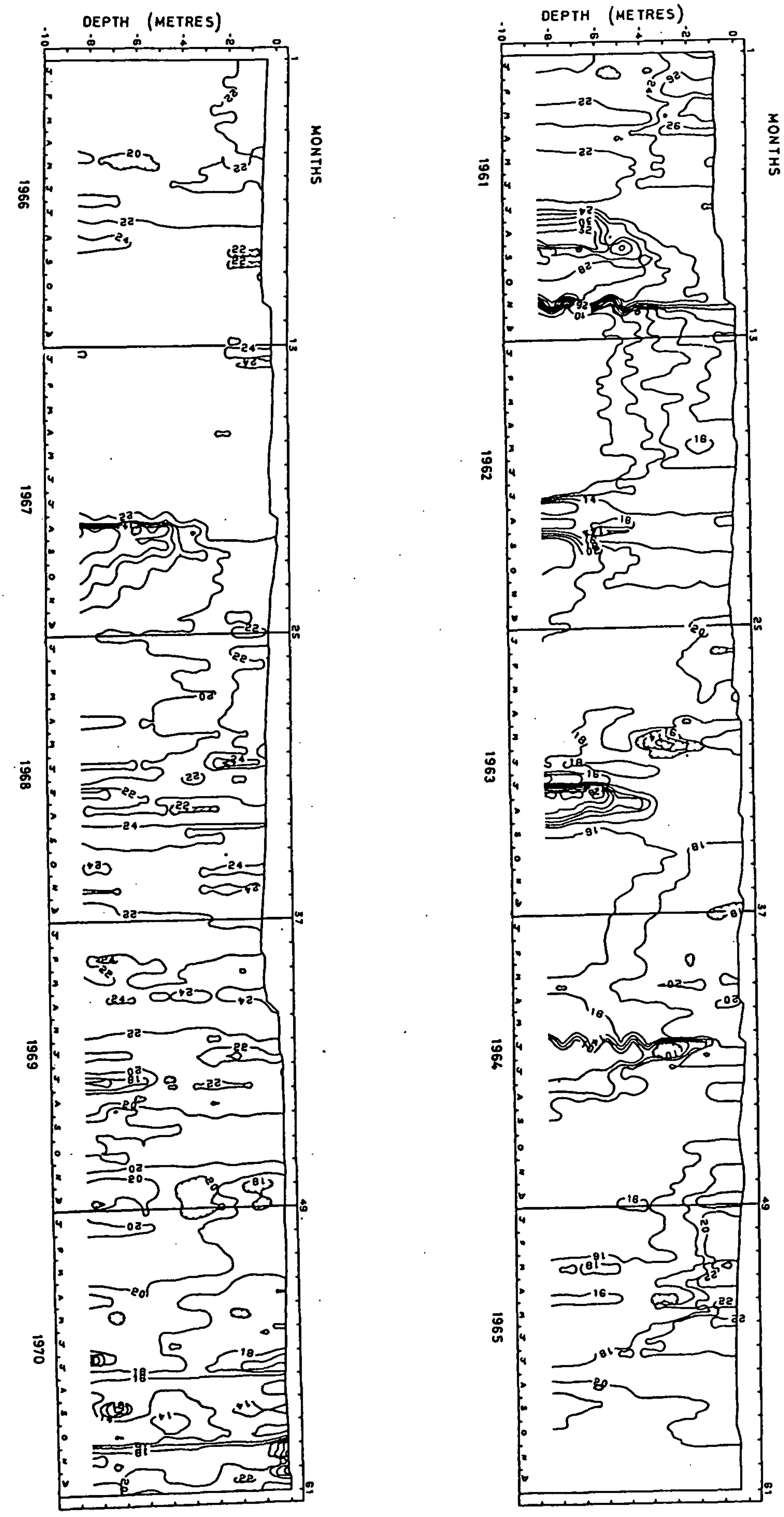

DEPTH (METRES)
MONTHS
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

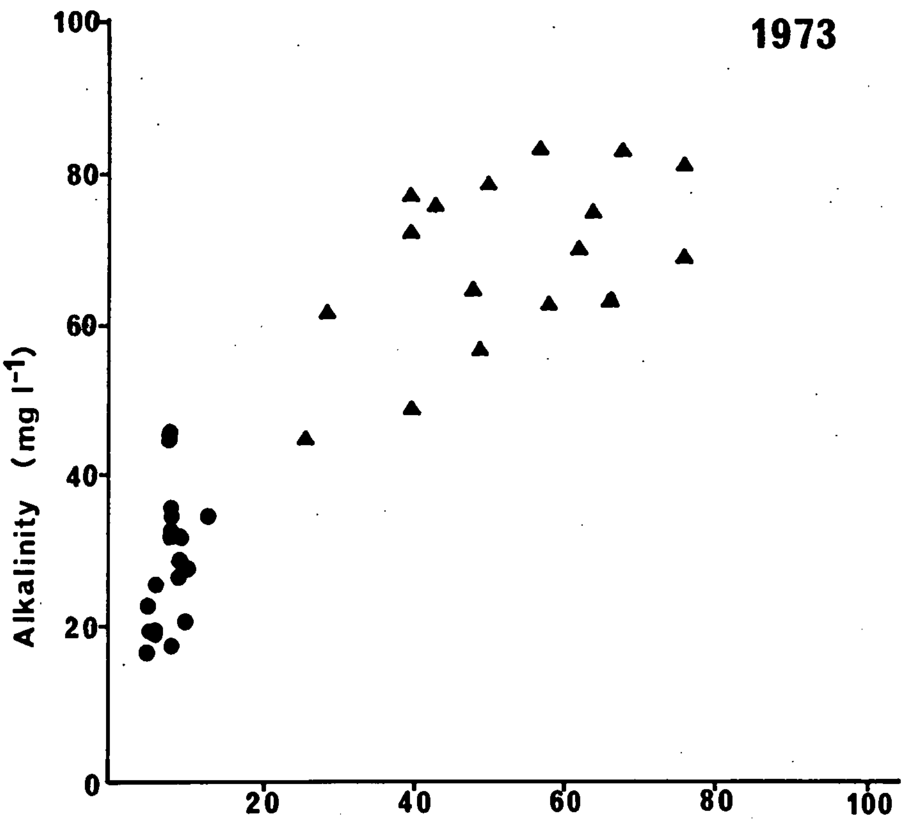
FIGURE 6.7

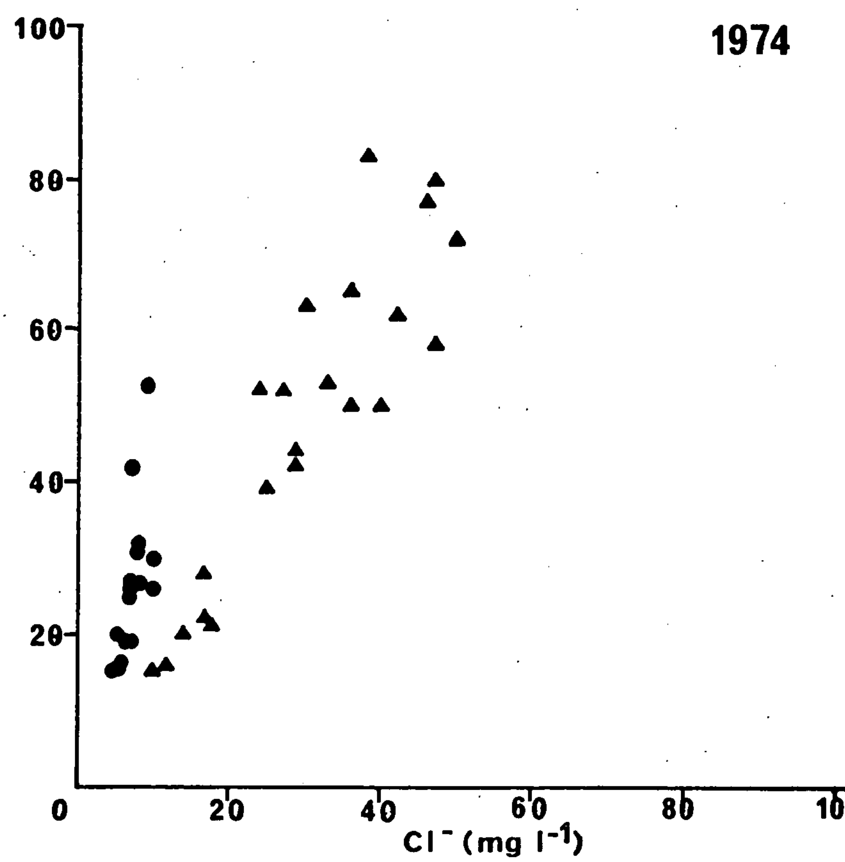
This figure consists of two pages, showing the scatter plots of Alkalinity (mg l^{-1} as $CaCO_3$) and Chloride concentration (mg l^{-1}) for the Wollondilly (triangles) and Cox Rivers (circles). The data are from samples collected at the inflow sample sites (Chapter 2, Fig. 2.1). Three years (1973, 1974, and 1979) are shown, which span almost the full range from very wet conditions (1974) to very dry conditions (1979). The plots demonstrate the chemical difference between these two major inflows, which probably arise from the different geology of the two catchments (see Introduction, Chapter 1). The greatest chemical divergence between the two rivers occurs in dry years, when the Wollondilly becomes more concentrated with respect to both chloride and alkalinity.

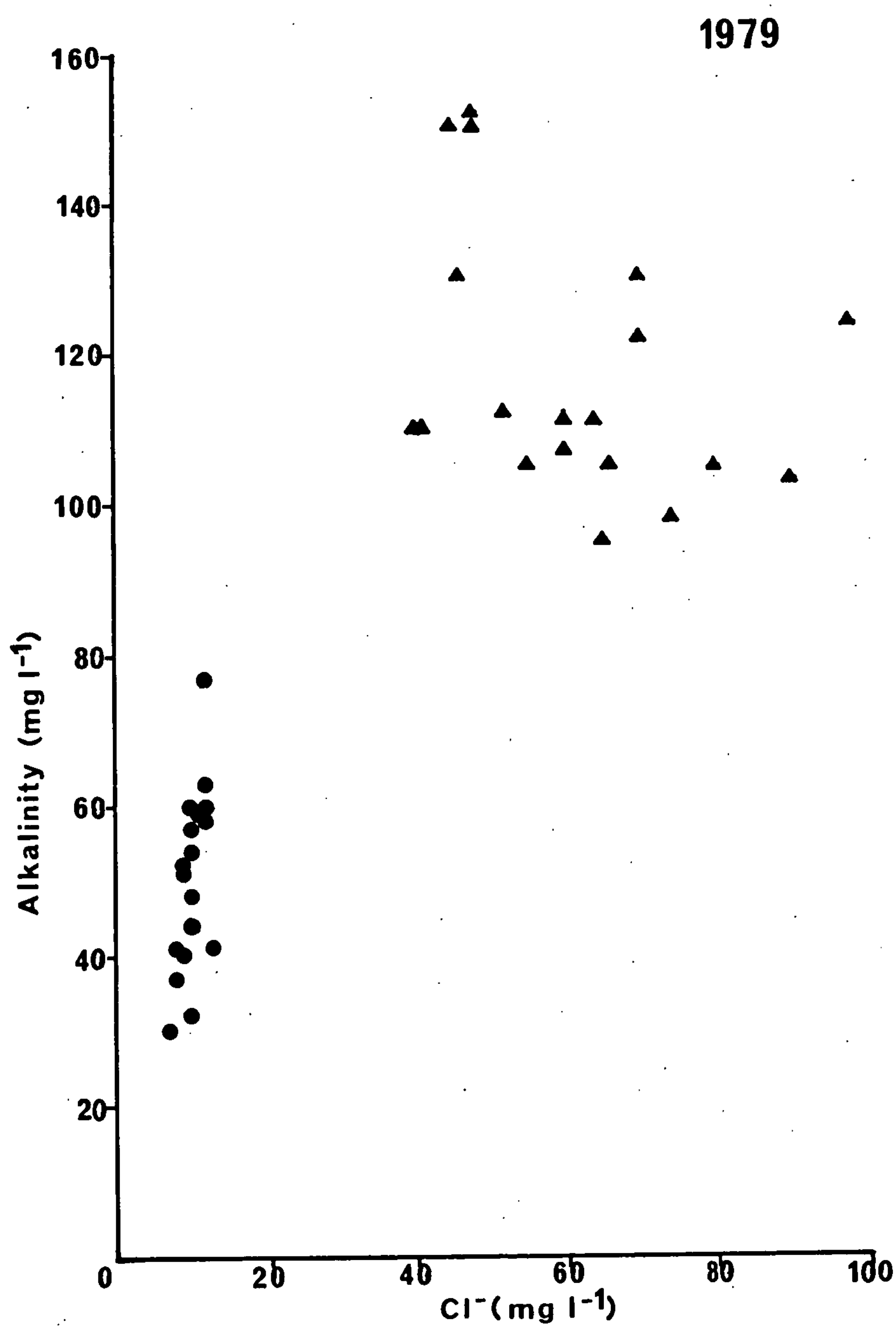
Symbols:-

Wollondilly River (▲)

Cox River (●)


1973
Alkalinity (mg l^{-1})
0
20
40
60
80
100


1974
Cl^{-} (mg l^{-1})
0
20
40
60
80
100


1979
Alkalinity (mg l-1)
160
140
120
100
80
60
40
20
0
20
40
60
80
100
Cl-(mg l-1)

one of the driest years in the period from 1962 - 1980.

The plots illustrate, firstly, that the Wollondilly River has generally higher chloride concentration and alkalinity than the Cox River. Secondly, that the Wollondilly River has more variable chloride concentration and alkalinity than the Cox River. Thirdly, it is evident that under widely varying flow regimes there is very little overlap of chloride concentration, in particular, within any one year. Finally, the plots indicate that the chloride concentration and alkalinity are reduced under high flow conditions in both rivers, although the Cox River is much less variable especially with respect to chloride concentration. There is insufficient data to know conditions during the "first flush" in either river.

The full range of chloride concentration recorded for the Wollondilly River was approximately 10 - 100 mg l^{-1}. In 1974 (wet) the range was c. 10 - 50 mg l^{-1}, and during 1979 (dry), values ranged from 40 - 100 mg l^{-1}. In contrast, the Cox River ranged from 5 - 15 mg l^{-1} overall.

Regression Analysis

This analysis relies on data from the Inflow Register (Chapter 3, Table 3.3), and data from direct flow gauging of the Wollondilly and Cox Rivers. Unfortunately, the Cox River flow data is unavailable for many of the larger inflows, which restricts the data to medium and low volume inflows (< 600 10^6 m^3). The data presented here has not been ln transformed.

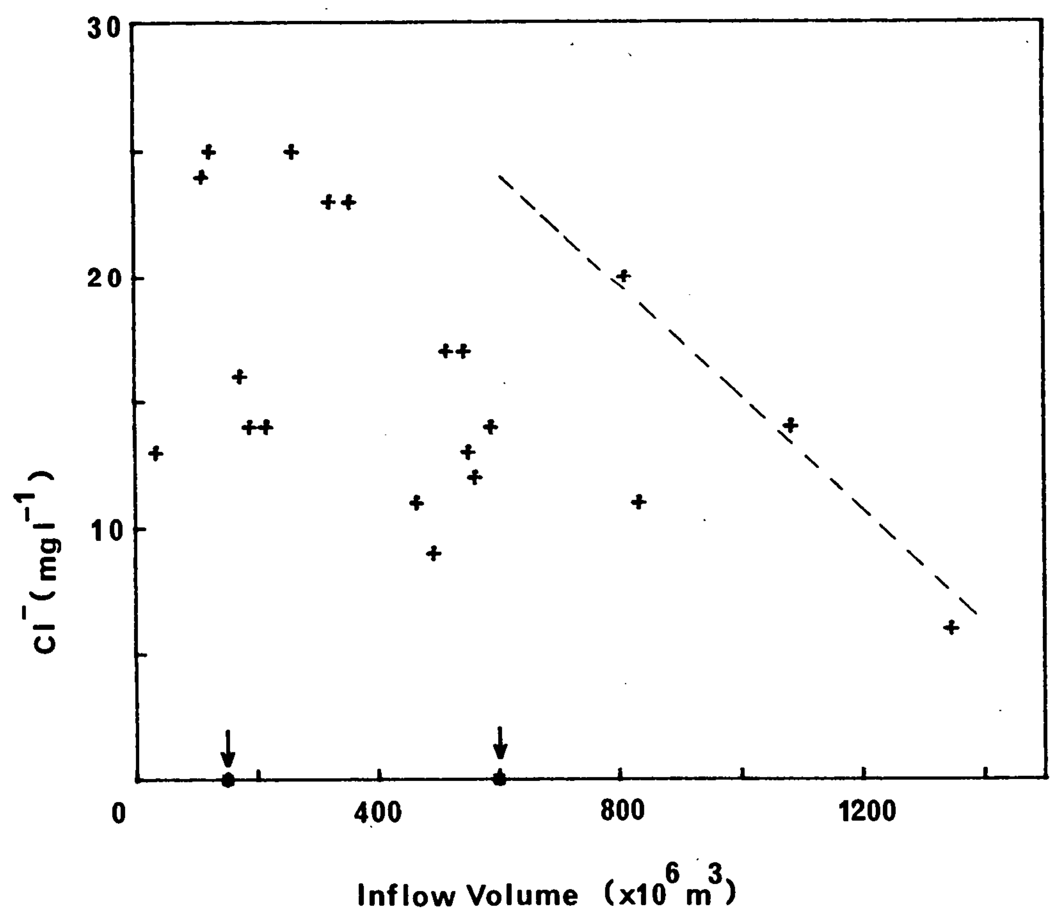
Two assertions may be made from data so far presented:-

1. Very high volume inflows from either river will probably cause a decrease in chloride concentration at site 3D.
2. The concentration of chloride at site 3D may be increased or lowered depending on the relative contribution of the two river systems to the total volume of inflow, particularly for lower volume inflows.

Fig. 6.8 shows a scatter plot of the chloride concentration against inflow

FIGURE 6.8

A scatter plot of the chloride concentration (mg l^{-1}) against inflow volume (10^6 m^3), for the 20 inflows from the December 1963 interflow to the June 1978 underflow, which resulted in a change of chloride concentration at site 3D (see Chapter 3, Table 3.3). Arrows mark the range of inflow for which the ratio of Wollondilly to Cox River inflow volumes is known and found to be significant in determining the chloride concentration of inflows. The broken line represents an approximate upper bound for chloride concentration of high volume inflows. The data tends to confirm the assertion that lower chloride concentration are usually associated with high volume inflows, and also indicates a greater scatter of chloride concentrations for lower volume inflows.



Cl⁻ (mg l⁻¹)
30
20
10
0
400
800
1200
Inflow Volume (x10⁶ m³)

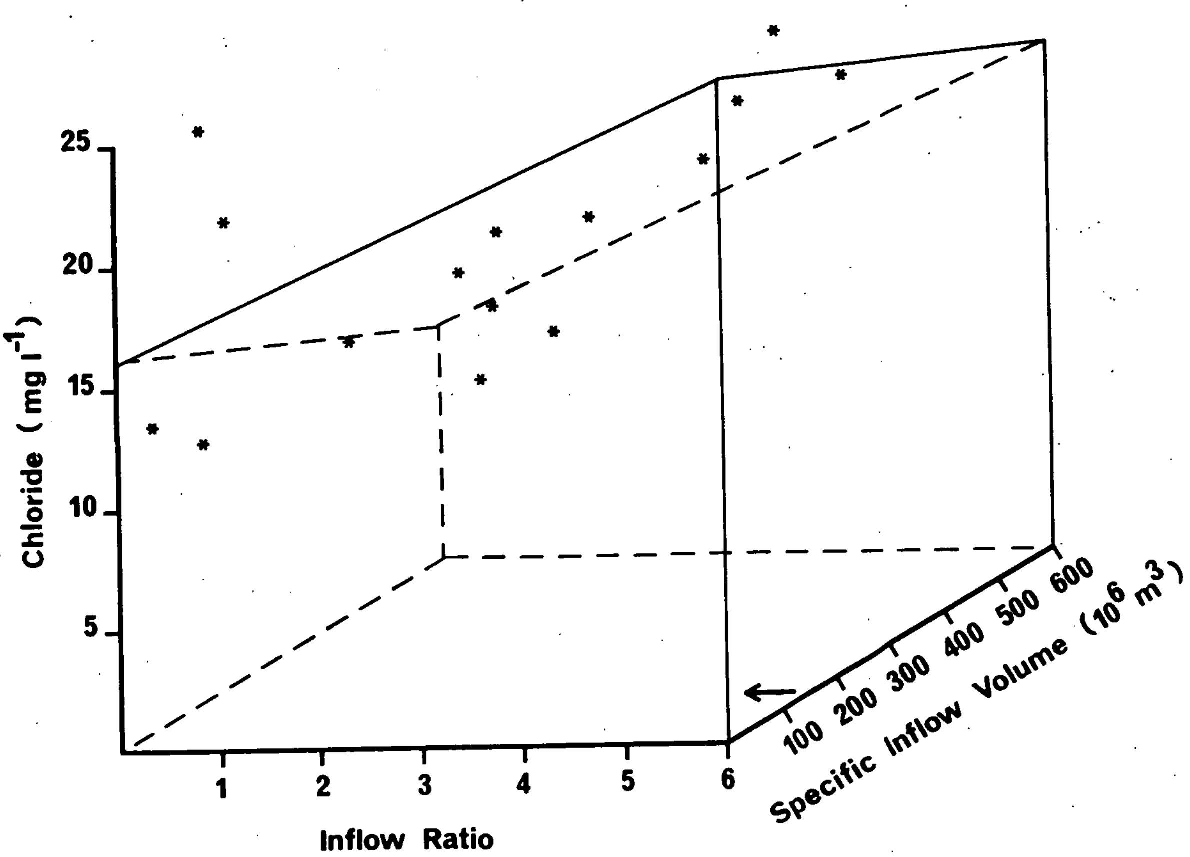
volume, for those 20 inflows from the December 1963 interflow to the June 1978 underflow, which resulted in a change of chloride concentration at site 3D. A linear regression of thesedata is just significant at the p = 0.05 (5%) level, explaining 27% of the variation in chloride concentration, and having a negative slope. Obviously, however, the relationship is not very close. Extension of this data set to include all 35 of the inflows for which inflow volumes are reported in Table 3.3 provides a more significant regression (at p = 0.01; 1%), but it still explains only 25% of the variation in chloride concentration and does not fit all the distributional requirements of the linear regression model. Nevertheless, the data tend to confirm the assertion that lower chloride concentration are usually associated with high volume inflows, and also indicate a greater scatter of chloride concentrations for lower volume inflows.

Only 15 inflows could be assigned a ratio of Wollondilly/Cox River inflow volume (directly gauged). The chloride concentration for these inflows is plotted against the total inflow volume (Table 3.3) and the inflow ratio in Fig. 6.9. The plane drawn on the diagram represents a multiple regression using all the data (n = 15). Again, a slight negative slope is indicated for the relationship between inflow volume and chloride concentration, but a more significant relationship (positive slope) is found between the inflow ratio and the chloride concentration. The scatter of points at low inflow volumes is also apparent. The multiple regression explains 52% of the variation in chloride concentration, but is only significant at p = 0.05. Further, the contribution of inflow volume is found to be insignificant, and the step-down regression is simply that of chloride concentration on inflow ratio, explaining 37% of the variation in chloride concentration and only significant at p = 0.05.

In view of the scatter of points at low inflow volumes, it is reasonable to exclude low volume inflows from the regression. Applying an arbitrary cut-off at 150 10^6 m^3 (marked with an arrow on Fig. 6.9) leaves 10 inflows

FIGURE 6.9

The chloride concentrations of 15 inflows, for which the ratio of Wollondilly to Cox River inflow volume (directly gauged) could be determined, are plotted against the corresponding total inflow volume (see the inflow register; Chapter 3, Table 3.3) and the inflow ratio. The plane drawn on the diagram represents a multiple regression using all the data (n = 15). There is a slightly negative relationship between inflow volume and chloride concentration, but a more significant relationship (positive slope) is found between the inflow ratio and the chloride concentration. The scatter of points at low inflow volumes is also apparent. An arrow marks the arbitrary cut-off point applied to the subsequent regression of chloride concentration on the inflow ratio (see Fig. 6.11).



25
20
15
10
5
Chloride (mg l-1)
1
2
3
4
5
6
Inflow Ratio
100 200 300 400 500 600
Specific Inflow Volume (10^6 m^3)

for which inflow ratio data exists. This includes all the available data from Table 3.3, regardless of whether the inflow resulted in a change of the ambient chloride concentration or not. For this unfortunately small data set, the inflow ratio explains 82% of the variation in chloride concentration. This regression is shown in Fig. 6.10, with 95% confidence and prediction limits. The resultant equation is:-

$$\begin{array}{llll} Cl^- = & 2.854\ Vol_{W/C} & + & 8.885 \quad (F = 35.5;\ p = 0.00034) \\ SE & 0.479 & & 1.394 \end{array}$$

Note that the data is in arithmetic units, because it accords with the requirements of the linear regression model without ln transformation. It may also be noted that y-intercept of this regression does not reflect the known average conditions at site 3D (c. 18 - 20 mg l^{-1} chloride concentration) in the absence of inflow. There is, in fact, no reason why it should since the regression can only be applied to the response of chloride concentration at site 3D to inflows between c. 150 and 600 10^6 m^3, and the projection of the inflow ratio to zero is not meaningful.

The analysis supports both of the assertions given earlier, although the role of the inflow ratio appears most significant for inflows of medium total volume. At volumes less than 150 10^6 m^3 neither the total inflow volume nor the ratio of Wollondilly to Cox River inflow are able to explain variation in chloride concentration at site 3D. For flows greater than 600 10^6 m^3 no ratio data exist.

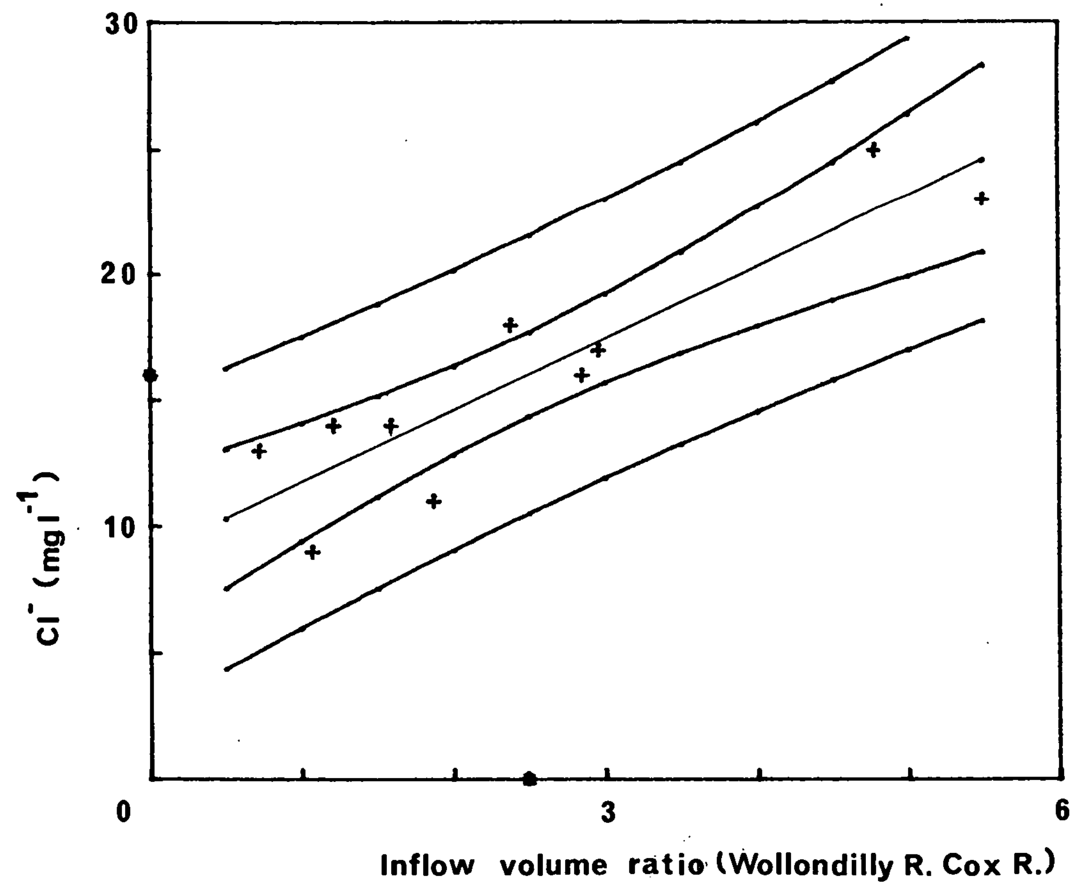
While, with the values experienced at site 3D, chloride is unlikely to be significant in water quality considerations, the above data indicate its utility as another means of recognizing the arrival of an inflow at 3D. It also offers some assessment of the relative contribution of the two major rivers to an inflow affecting the water column at site 3D. Finally, there is some indication that consideration of the total inflow volume, taking into account

FIGURE 6.10

The regression of chloride concentration (mg l^{-1}) on the ratio of Wollondilly and Cox River inflow volumes (dimensionless) for 10 inflows from Table 3.3. An arbitrary cut-off at 150 10^6 m^3 (marked with an arrow on Fig. 6.9) is applied. The data is in arithmetic units, because it accords with the requirements of the linear regression model without ln transformation. The inflow ratio explains 82% of the variation in chloride concentration (n = 10). The 95% confidence (inner) and prediction (outer) limits are marked on the plot.

$$Cl^- = 2.854\ Vol_{W/C} + 8.885 \quad (F = 35.5;\ p = 0.00034)$$

SE 0.479 1.394



30
20
10
0
3
6
Cl⁻ (mg l⁻¹)
Inflow volume ratio (Wollondilly R. Cox R.)

the relative contribution of the Wollondilly and Cox Rivers, may also provide information about the change in alkalinity at site 3D in response to inflows. This could be of some use to management, because the efficiency of chemical flocculation can be affected by low alkalinity (Linsley and Franzini 1972).

CHAPTER 7

INFLOW
AND
PHYTOPLANKTON
BIOMASS

INTRODUCTION

At site 3D, Lake Burragorang is characterised by comparatively low concentration of chlorophyll-a (used here to indicate phytoplankton biomass). Concentrations of less than 5 mg m^{-3} are common, and the annual maximum record, since 1970, has rarely exceeded 25 - 30 mg m^{-3}. Under favourable conditions the concentration of chlorophyll-a may exceed these values by about one order of magnitude in another Australian water supply reservoir, Mount Bold Reservoir, South Australia (Ferris 1977; reported by Ganf 1980a). It is therefore apparent that various limitations exist which curtail algal growth in Lake Burragorang. Well recognised amongst these limiting factors, are the availability of nutrients, particularly phosphorus (Schindler 1978), and light (Talling 1971). Other significant limiting factors include lake retention time and various indirect effects of lake morphology, such as depth and the degree to which the lake surface is exposed to wind. The major limiting factors differ from lake to lake, and there may be periodic shifts in the relative importance of these factors within the one lake.

In Lake Burragorang inflow has an important effect on several of the major factors that may limit algal growth. The suspended particulate matter, associated with major inflows, affects light penetration and also serves as a vehicle for the transport of nutrients into the lake.

TURBIDITY AND SECCHI DEPTH

Data for the analysis come from records for the period 1961 to 1978, and data pairs are limited to those where surface turbidity exceeds 3 Hellige units, to avoid the situation where a very large percentage of the data occupies a very small part of the total data range.

Analysis of surface turbidity data with closely matched data for Secchi disc depth (metres) indicates a highly significant linear relationship between the two parameters, explaining 82% of the variation in the dependent variate ($n = 150$; $r^2 = 0.82$). Two equations are given, the first has surface turbidity as the independent variate, and the second is a reversal of the first, having reciprocal Secchi depth as the independent variate:-

$\ln(1/\text{Secchi}) =$	$0.867 \ln \text{Turb}_S$	$- 2.057$	$(F = 697, p = 7.4\ 10^{-58})$
SE	0.033	0.069	

$\ln \text{Turb}_S =$	$0.952 \ln(1/\text{Secchi})$	$+ 2.305$	$(F = 697, p = 7.4\ 10^{-58})$
SE	0.036	0.028	

Fig. 7.1 shows the second regression, with 95% confidence and predictive limits.

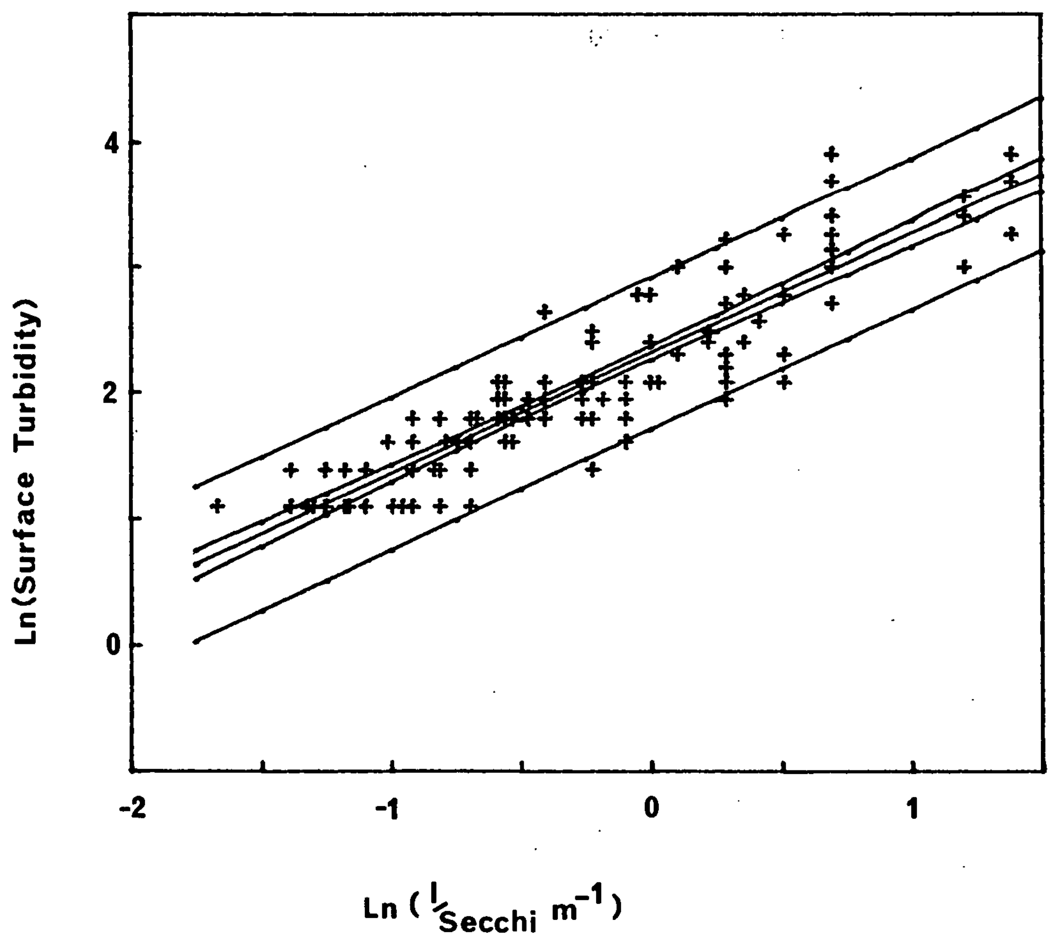
It should be noted that the data set does not quite fulfil the requirements of the linear regression model, though the ln transformation of both variates considerably improves the distribution of residuals. In both cases the absolute values of the residuals are found to be correlated with the independent variate. According to Sokal and Rohlf (1981) this casts doubt on the use of the relationships for predictive purposes, but does not alter the fact of a close relationship between turbidity and Secchi depth. In my opinion, the finding of inappropriate distribution of the residuals probably rests on relatively few outliers, such as the points in Fig. 7.1 that lie outside the 95% predictive limits, and may not seriously challenge the use of the regression. Alternatively, the relationship between the two variates may be better approximated by a quadratic function. I have not investigated this possibility here.

FIGURE 7.1

The figure shows a close relationship between surface turbidity (Hellige units) and Secchi transparency (m), for closely matched data excluding occasions when turbidity was less than 3 units. The use of reciprocal Secchi depth gives a positive slope. The regression explains 82% of the variation in the dependent variate (n = 150; r^2 = 0.82). The data set does not quite fulfil the requirements of the linear regression model, though the ln transformation of both variates considerably improves the distribution of residuals. The 95% confidence limits fall very close to the regression line, while the 95% prediction limits indicate the extra uncertainty involved in prediction of turbidity from single measurements of Secchi depth.

$$\ln Turb_S = 0.952 \ln(1/Secchi) + 2.305 \quad (F = 697,\ p = 7.4\ 10^{-58})$$

SE	0.036	0.028


Ln(Surface Turbidity)
4
2
0
-2
-1
0
1
Ln (1/Secchi m-1)

Secchi and Euphotic Depths

Recently, the acquisition of a quanta sensor (by the MWS&DB) has enabled the direct measurement of light extinction in Lake Burragorang. Using a small data set covering one year (between 1979 and 1980) the relationship between Secchi and euphotic depth is examined. Secchi disc transparency is a poor estimator of euphotic depth for Lake Burragorang, under the comparatively clear water conditions that prevailed (extinction coefficients in ln units, for downwelling irradiance of wavelength 400 - 700 nm, ranged from 0.27 - 0.59 m^{-1}).

Quite close relationships between Secchi and euphotic depths have been reported by other workers (i.e. Walker et al 1983) and Secchi transparency has been widely used as an index of euphotic depth, although the multiplicative conversion from Secchi depth to euphotic depth is variously reported as less than 1.0 (King pers. comm.) for clear humic lakes to at least 2.0 (see Carlson 1977).

TOTAL PHOSPHORUS AND CHLOROPHYLL

This section contains the text of a paper recently published in the Australian Journal of Marine and Freshwater Research:-

Ferris, J. M., and Tyler, P. A. (1985). Chlorophyll-total phosphorus relationships in Lake Burragorang, New South Wales, and some other Southern Hemispere lakes. *Aust. J. Mar. Freshw. Res.* **36**,157-68.

Apart from removing the abstract and placing the materials and methods in Chapter 2, the text is little altered from the paper. There is a small amount of repetition of material covered in the introduction (Chapter 1). Since the paper was submitted, a little more information has become available concerning phosphorus-chlorophyll relationships in Southern Hemisphere lakes. Pridmore et al (1985) examine some lakes in the North Island of New Zealand. This study provides further confirmation of the general concordance of Southern and Northern Hemisphere lakes with respect to phosphorus-chlorophyll relationships.

Introduction

Widespread concern to prevent or abate eutrophication prompted many studies on the role of phosphorus, and that chlorophyll-a concentrations in the euphotic zone are frequently determined by phosphorus concentrations has been amply demonstrated (Smith and Shapiro 1981). The extensive literature for individual European and North American lakes fuelled second generation studies, formulating empirical relationships from data for numerous lakes (Dillon and Rigler 1974; Jones and Bachman 1976; Schindler 1978; OECD 1982), predicting consequences of given nutrient regimes or indicating steps necessary to maintain or improve water quality. Schindler (1978), reviewing I.B.P. data, noted a heavy bias towards data for Northern hemisphere lakes, stating that data for the Southern Hemisphere were "sorely needed" to permit a true global analysis. The same bias is evident in the recent OECD (1982) study. In this paper, we investigate relationships between chlorophyll and total phosphorus concentrations, over 11 years, in a deep impoundment, Lake Burragorang, New South Wales, and re-analyse data from tropical Australia, New Zealand, and South Africa to appraise their degree of conformity with prevailing Northern Hemisphere concepts.

Study Area

Lake Burragorang, formed in 1960 behind Warragamba Dam, is a major water supply for Sydney. It is dendritic, Y-shaped, with the Cox arm receiving treated sewage via Kedumba Creek (Chapter 2, Fig. 2.1). Public access to the catchment is restricted closer than 3 km to full supply level. The lake is warm-monomictic, with short circulation periods, and has relatively high transparency. Some basic features of the reservoir and its relatively undeveloped catchment are given in Chapter 1, Table 1.1. Further data are reported by Jolly (1966b) and Bowen and Smalls (1980). In view of Lake Burragorang's size and morphometric complexity, it may be expected to show considerable variation between sample sites. The extent of this site to site variation is not investigated here, though there is some individual consideration of the site adjacent to Warragamba Dam. The treatment in this study represents an averaging of a spatially and temporally variable system.

Materials and Methods

The methods appropriate to this section are given in Materials and Methods, Chapter 2.

Results

When gravimetric ratios of total nitrogen (tN) concentration to total phosphorus (TP) concentration were calculated for the offtake, at Warragamba Dam, annual means ranged from 23 to 48, with a 6 year mean of 34. By the criteria of either Sakamoto (1966) or Smith (1982a), Lake Burragorang is substantially a phosphorus-limited system, approaching the point where the influence of nitrogen is negligible. Note that the ratios (tN:TP) presented here will underestimate the actual TN:TP ratios. That light limitation is unlikely to be significant in Lake Burragorang is demonstrated by the fact that the geometric mean annual Secchi transparency, for site 3D (1970 - 1980), was 4.0 m. This compares with the geometric mean annual Secchi transparency of 3.3 m for the lakes (n = 94) in the eutrophication study conducted by the OECD (1982).

Table 7.1 contains regressions from Southern Hemisphere lakes and reservoirs for which published TN:TP ratios suggest phosphorus limitation. For Lake Burragorang, equation (1) is derived from a regression of annual maximum chlorophyll-a concentration on annual mean total phosphorus concentration, following Schindler (1978) and OECD (1982). Equations (4) - (7) are for data from New Zealand, South Africa and the Northern Territory, Australia. Table 7.2 shows some published chlorophyll-phosphorus regressions for the Northern Hemisphere. The regression slopes vary from 0.83 to 1.58. The slopes of Lake Burragorang regressions are similar to those reported elsewhere (Tables 7.1 and 7.2) and within the range (0.4 - 2.2) for individual lakes in the Northern Hemisphere (Smith and Shapiro 1981).

Fig. 7.2 contains graphs of the Lake Burragorang regressions, with 95% confidence intervals and 95% prediction limits for pooled data from all three sample sites. Fig. 7.2 (upper panel) depicts annual maximum chlorophyll-a concentration regressed on annual mean total phosphorus concentration. The even spacing of points along the regression line is statistically satisfactory. Site to site variation in the cocncentrations of phosphorus and chlorophyll-a are evident from the distribution of points in three overlapping groups. The numerical order of group mean total phosphorus concentration is site 3D (outlet) < Cox River above Kedumba Creek (inlet) < Kedumba Creek (sewage enriched). The 95% prediction limits are broad, as reported elsewhere (OECD 1982). A separate regression line for site 3D alone is included. Fig. 7.2 (lower panel) shows the regression of annual mean chlorophyll-a concentration on annual mean total phosphorus concentration, described by equation (3).

From equations (1) and (2) (Table 7.1), a nominal annual mean total phosphorus concentration of 20 mg m^{-3} predicts a corrected annual maximum chlorophyll concentration of 18 mg m^{-3} for pooled data and 32 mg m^{-3} for site 3D. Equation (3) may be directly compared with that of Schindler (1978), in that both use annual mean concentrations of total phosphorus and

Table 7.1 Chlorophyll–phosphorus regression equations for selected Southern Hemisphere lakes

C, Chlorophyll a concentration (mg m^{-3}); P, total phosphorus concentration (mg m^{-3}). F, multiplicative correction factor. A bar over the symbol for chlorophyll or phosphorus indicates a mean value; the superscript g indicates a geometric mean. The following subscripts are used: a, annual; s, surface values; m, maximum; sp, spring; su, summer

Reference	Equation	n	r^2	F	Location
	Individual regressions				
(1) This study, all sites	$\ln C_{a,m} = 1 \cdot 502 \ln \overline{P}_a - 1 \cdot 754$	33	0·60	1·161	Lake Burragorang, N.S.W.
(2) This study, site 3D	$\ln C_{a,m} = 2 \cdot 191 \ln \overline{P}_a - 3 \cdot 197$	11	0·77	1·099	Lake Burragorang, N.S.W.
(3) This study, all sites	$\ln \overline{C}_a = 1 \cdot 130 \ln \overline{P}_a - 1 \cdot 806$	33	0·67	1·066	Lake Burragorang, N.S.W.
	Regional regressions				
(4) McColl (1972)	$\ln \overline{C}_a^g = 0 \cdot 984 \ln \overline{P}_a^g - 1 \cdot 767$	7	0·85	1·096	New Zealand
(5) Walmsley and Butty (1980)	$\ln C_{a,m} = 1 \cdot 153 \ln \overline{P}_a - 1 \cdot 302$	10	0·71	1·179	South Africa
(6) Walmsley and Butty (1980)	$\ln \overline{C}_a = 0 \cdot 837 \ln \overline{P}_a - 1 \cdot 089$	10	0·74	1·078	South Africa
(7) Walker and Tyler (1983)[A]	$\ln C_s = 1 \cdot 258 \ln P_s - 2 \cdot 045$	48	0·53	1·196	Northern Territory

[A] Walker and Tyler used individual samples taken during the 7–8-month dry season, not means.

Table 7.2 **Chlorophyll–phosphorus regressions for Northern Hemisphere lakes**

Slopes of $\log_{10}$ equations do not change on transformation to ln units, but y-intercepts have been transformed in the sequence $\log_{10} \rightarrow$ arithmetic $\rightarrow$ ln. Symbols, subscripts and superscripts are as for Table 7·1

Reference	Equation	n	r^2	Location
	Individual regressions			
(8) Smith and Shapiro (1981)[A]	$\ln \overline{C}_{su} = 1 \cdot 20 \ln \overline{P}_{June\text{-}Sept} - 1 \cdot 27$	16	0·93	Lake Washington, U.S.A.
	Regional regressions			
Sakamoto (1966)	$\ln \overline{C}_{su} = 1 \cdot 583 \ln P_{sp} - 2 \cdot 611$	31	0·95	Japan
Carlson (1977)	$\ln \overline{C}_{su} = 1 \cdot 449 \ln \overline{P}_{su} - 2 \cdot 442$	43	0·72	U.S.A.[B]
Hickman (1980)	$\ln C_{su.m} = 0 \cdot 828 \ln P_{sp} - 1 \cdot 094$	17	0·28	Canada
Prepas and Trew (1983)	$\ln \overline{C}_{su} = 1 \cdot 119 \ln P_{sp} - 1 \cdot 557$	29	0·67	Canada[C]
Prepas and Trew (1983)	$\ln \overline{C}_{su} = 1 \cdot 573 \ln P_{sp} - 3 \cdot 056$	24	0·63	Canada[C]
Prepas and Trew (1983)	$\ln \overline{C}_{su} = 1 \cdot 146 \ln \overline{P}_{su} - 1 \cdot 522$	34	0·81	Canada[C]
	Worldwide regressions			
Dillon and Rigler (1974)	$\ln \overline{C}_{su} = 1 \cdot 449 \ln P_{sp} - 2 \cdot 616$	46	0·90	
Jones and Bachmann (1976)	$\ln \overline{C}_{su} = 1 \cdot 46 \ln \overline{P}_{su} - 2 \cdot 510$	143	0·90	
Schindler (1978)[D]	$\ln \overline{C}_{a} = 1 \cdot 213 \ln \overline{P}_{a} - 1 \cdot 953$	~80	0·77	

[A] The correction factor for this equation is 1·021.
[B] Carlson does not specify the source of all his data, but that which he does specify is confined to North America.
[C] In the second of these three equations, saline lakes were excluded.
[D] Two sites from the Southern Hemisphere are included in Schindler's data set.

FIGURE 7.2

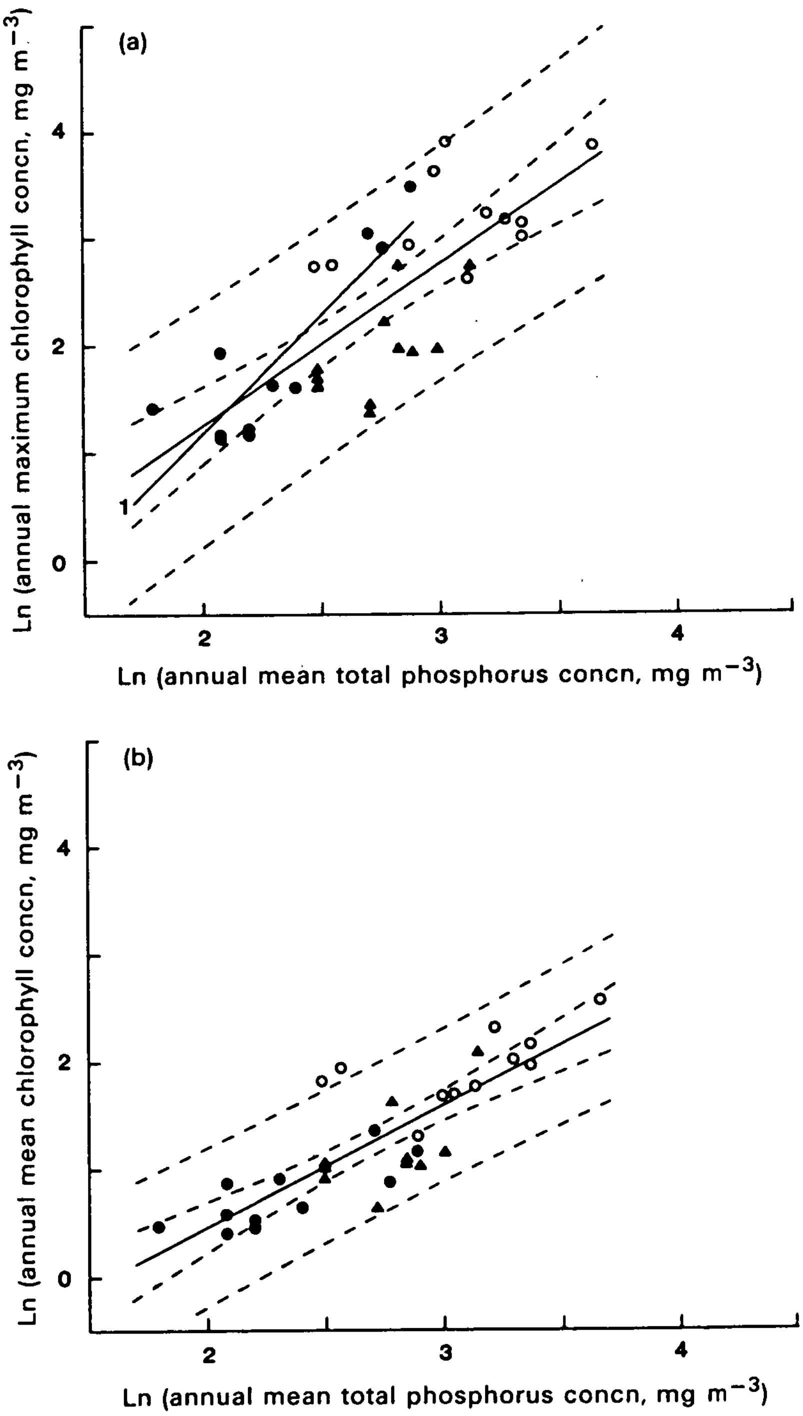
Chlorophyll-phosphorus regressions for Lake Burragorang, 1970 - 1980. The 95% confidence (inner lines) and 95% prediction limits are marked with broken lines. All variates have been ln transformed. The upper panel shows annual maximum chlorophyll-a concentration (mg m^{-3}) regressed on annual mean total phosphorus (mg m^{-3}). The numbered line (1) is a regression for site 3D only. The un-numbered line is for pooled data from all three sample sites. The lower panel shows annual mean chlorophyll-a regressed on annual mean total phosphorus. Equations for these regressions are given in Table 7.1.

Symbols:-

Site 3D (●)

Kedumba Creek (o)

Cox River above Kedumba Creek (▲)



(a)
Ln (annual maximum chlorophyll concn, mg m^{-3})
Ln (annual mean total phosphorus concn, mg m^{-3})
0
2
4
2
3
4
1
(b)
Ln (annual mean chlorophyll concn, mg m^{-3})
Ln (annual mean total phosphorus concn, mg m^{-3})
0
2
4
2
3
4

chlorophyll-a. For an annual mean total phosphorus concentration of 20 mg m^{-3}, both equations predict an annual mean chlorophyll-a concentration of 5 mg m^{-3} (predicted chlorophyll-a concentration, in both cases, is uncorrected as F was not reported for Schindler's equation). The more recent OECD (1982) equation also predicts an annual mean chlorophyll-a concentration of 5 mg m^{-3} for an annual mean total phosphorus concentration of 20 mg m^{-3}. If a nominal value of 40 mg m^{-3} is chosen for the annual mean total phosphorus concentration (the upper limit of our data), then the predicted annual mean chlorophyll-a values are 10 mg m^{-3} (OECD 1982), 11 mg m^{-3} (Lake Burragorang) and 12 mg m^{-3} (Schindler 1978). It is evident that our equation is in close agreement with the global models of Schindler (1978) and the OECD (1982).

We feel that the utility of chlorophyll-phosphorus regressions lies in their prediction limits in arithmetic terms, and in Table 7.3 these are derived for chosen concentrations of of total phosphorus, from our equations for Lake Burragorang, and for re-analysed data for the Southern Hemisphere. Lake Washington was chosen as a Northern Hemisphere example because it has been studied for at least as long as Lake Burragorang.

Discussion

The slopes of chlorophyll-phosphorus regressions vary considerably, the range being greater for individual lakes than for regional or world-wide studies (Table 7.2). The greatest variation we know (0.4 - 2.2) is between regressions for a series of individual lakes studied by Smith and Shapiro (1981). Several authors have examined the reasons for thr overall variation in chlorophyll-phosphorus regressions (Nicholls and Dillon 1978; Canfield and Bachmann 1981; Smith 1982a; Hoyer and Jones 1983). The slopes of the two regressions for Lake Burragorang (pooled data: 1.13 and 1.50) are well within world experience. Combined with the other Southern Hemisphere studies,

Table 7.3 **Arithmetic ranges of chlorophyll concentrations within 95% prediction limits for selected concentrations of total phosphorus**

Predictions are from numbered regression equations (see Tables 7·1&7·2) for Lake Burragorang, other Southern Hemisphere sites, and Lake Washington. The upper prediction at each total phosphorus concentration is given first

Regression	Chlorophyll concn (mg m^{-3}) at total phosphorus concn (mg m^{-3}) of: 5	10	15	20	25	30	40	50	75
(1)	6·4	16·5	29·9	46·4	65·8	88·1	141·0		
	0·6	1·8	3·4	5·2	7·2	9·3	13·8		
(2)	4·7	17·7	45·6	95·7					
	0·4	2·3	5·2	8·8					
(3)	2·2	4·6	7·1	9·9	12·9	16·0	22·7		
	0·5	1·1	1·7	2·4	3·0	3·7	5·0		
(4)	3·2	5·5	8·0	10·6	13·3	16·1	22·2	28·6	
	0·2	0·5	0·8	1·0	1·2	1·5	1·9	2·2	
(5)	10·5	18·6	27·0	35·8	45·3	55·4	77·2	101·6	172·1
	0·3	0·8	1·4	2·1	2·7	3·4	4·7	6·0	9·1
(6)	4·3	6·6	8·7	10·8	12·8	14·8	18·8	23·0	33·5
	0·4	0·8	1·2	1·6	1·9	2·3	2·9	3·4	4·6
(7)	3·8	8·3	13·2	18·6	24·4	30·6	43·9	58·4	99·5
	0·3	0·7	1·2	1·7	2·3	2·9	4·1	5·4	8·8
(8)	3·2	7·2	11·5	16·1	21·0	26·2	37·4	49·4	
	1·1	2·8	4·6	6·5	8·5	10·5	14·7	19·1	

which include regressions for a wide range of lake types (and differences in regression subvariables; Table 7.1), the range of regression slopes (0.84 - 1.50) matches that for regional studies of the Northern Hemisphere (Table 7.2).

A regression line is exactly defined, not simply by its slope but by its slope and intercept (Sokal and Rohlf 1981). Intercepts have been compared by substituting chlorophyll concentration as the independent variate. The y-intercept of these reversed regressions then gives total phosphorus concentrations corresponding to 1 mg m^{-3} chlorophyll-a (because the origin of double-ln plots is 1:1 in arithmetic units). For regressions quoted in Table 7.1, intercepts range from 5 to 13 mg m^{-3} total phosphorus, with only one value, that of Walker and Tyler (1983), greater than 8. The range for Lake Burragorang is 5 - 7 mg m^{-3}. For Lake Washington (Smith and Shapiro 1981) the value is 3 mg m^{-3}. On the basis of slope and, it seems, intercept, Southern Hemisphere regressions reported here show substantially the same relationship between chlorophyll and phosphorus as in the Northern Hemisphere. However, two idiosyncracies of the Southern Hemishere should be noted.

Firstly, in two studies, satisfactory regressions were obtained only if turbid lakes were excluded. Walmsley and Butty (1980) excluded reservoirs with annual mean Secchi depths $\leq$ 0.4 m, and Walker and Tyler (1983) excluded billabongs with turbidity > 10 NTU (equivalent to transparencies less than 1.3 m). The OECD (1982) applied a similar rationale in screening their data, rejecting one of two Australian examples. This limitation is especially significant in Australia, with so many turbid waters (Kirk 1977a, 1977b; Ganf 1980b; Williams 1983). Canfield and Bachmann (1981), comparing natural and artificial lakes in the United States, suggested that non-algal turbidity may significantly modify chlorophyll-phosphorus relationships in artificial lakes. Jones and Novak (1981) found that, in the Lake of Ozarks, U.S.A., high inorganic turbidities were associated with chlorophyll-a concentrations that were from 9 to 36% of those expected from the concentrations of total phosphorus. In the few studies that have taken account of the non-algal turbidity, two distinct strategies have been adopted. A threshold approach is inherent in the data selection used by Walmsley and Butty (1980), the OECD (1982), and Walker and Tyler (1983, 1984). Walmsey and Butty (1980) provided a separate multiple regression equation to predict the concentration of chlorophyll-a in South African reservoirs with mean annual Secchi depths ≤ 0.8 m. In contrast, Hoyer and Jones (1983) accounted for the effects of non-algal turbidity over the full range of their data, by using the ratio of concentrations of inorganic suspended solids to total phosphorus in a general multiple regression. In a similarly general approach, Verduin et al (1978) and Rosich (1983) used the ratio of euphotic depth to mean mixed depth to modify the basic chlorophyll-phosphorus relationship and to account for the effect of non-algal turbidity. Smith (1982b) mentions two aspects of this effect: firstly, an increase in particle-associated "biologically unavailable" phosphorus and, secondly, a less favourable underwater light climate, both tending to reduce the concentration of chlorophyll-a relative to the total phosphorus concentration. Whether or not a single equation can adequately characterise the effect of non-algal turbidity on chlorophyll-a concentrations, for the full range of turbidity encountered in Australia and South Africa, remains to be determined. One of the more turbid sites included in Hoyer and Jones' (1983) study (Ozark arm of the Lake of Ozarks), which yielded annual means of turbidity and Secchi depth of 36 JTU and 0.4 m, respectively (Jones and Novak 1981), was considerably less turbid than three of the South African impoundments studied by Walmsley and Butty (1980), where annual means of turbidity and Secchi depth ranged from 53 to 76 JTU and from 0.10 to 0.17 m respectively.

The second departure from many previous studies is our use of annual maximum, as opposed to "growing season" means, of chlorophyll-a for Lake Burragorang and in re-analysing the data of Walmsley and Butty (1980). We chose to use the maximum annual record of chlorophyll-a for several reasons. Firstly, seasonal contrasts in Sydney's climate are small, and the growing season is considerable. In 11 years, the annual maximum chlorophyll-a concentration was recorded at any time between August and May (10 months). In contrast, Smith and Shapiro's (1981) "growing season" was 2 - 6 months. Secondly, inflow to Lake Burragorang is unpredictable, both seasonally and from year to year. Nominal retention times (lake volume/annual inflow) ranged from 0.5 to > 20 years between 1970 and 1980. Large inflows may flush out plankton at any time. Then, high phosphorus concentrations can coincide with a minimum of plankton. We believe the annual chlorophyll-a maximum better portrays the productivity of Lake Burragorang. Schindler (1978) mentions conceptual difficulties in using growing season means in a world-wide context, especially for tropical lakes, and the OECD (1982) uses both annual mean chlorophyll-a and annual maximum chlorophyll-a concentrations. Finally, we agree with Forsberg and Ryding (1980) that prediction of annual maximum chlorophyll-a concentration may be of paramount interest in practice.

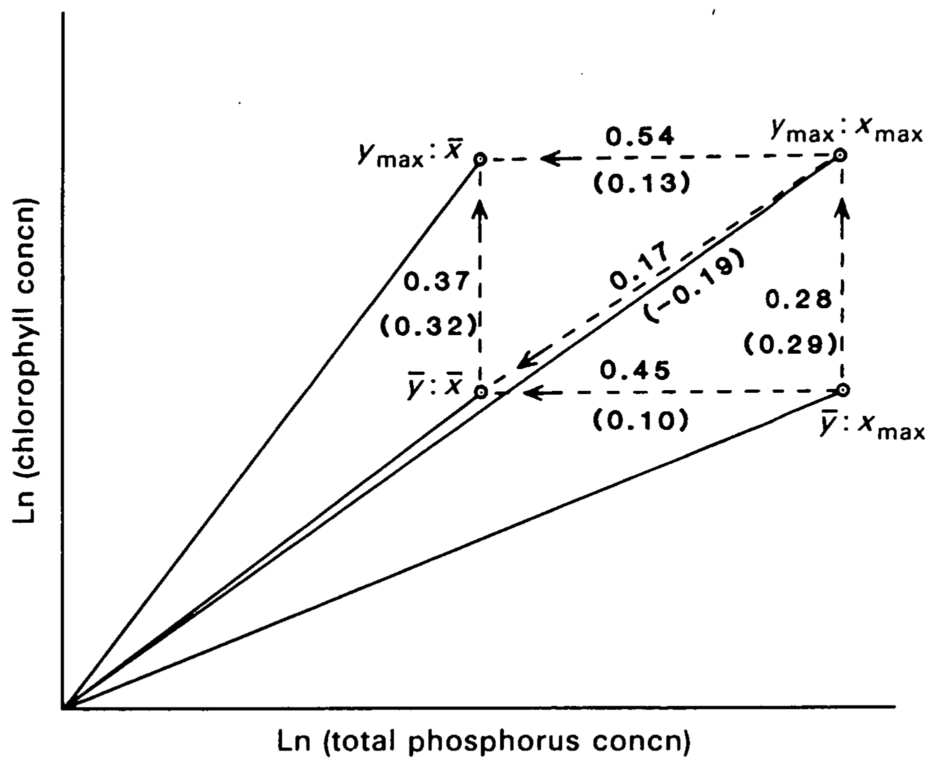
Some of the variation in slope of regressions reported in Tables 7.1 and 7.2 stems from the choice of different subvariables (e.g. annual mean, annual maximum, growing season mean, etc.) rather than from fundamental differences in the relationship between the primary variables, chlorophyll and phosphorus concentrations. Nicholls and Dillon (1978) observed that this source of variation was, in most cases, not evaluated due to a lack of adequate data. To estimate this variation, we measured the change in regression slope that could be generated by manipulation of the regression subvariables (annual mean and annual maximum) within our data set for Lake Burragorang and the data for South African reservoirs (Walmsley and Butty

1980). This analysis is schematically illustrated in Fig. 7.3 For these two data sets, one of an individual lake, the other of a regional study, the changes in slope range from 0.10 to 0.54, equivalent to as much as 70% of the variation of slope for regressions using similar combinations of regression subvariables (Table 7.2).

Chlorophyll-phosphorus regressions are widely used predictively in water-quality management and eutrophication abatement programs. We believe that the prediction of annual maximum chlorophyll-a concentration is of most practical significance, as short-lived algal blooms, which subsequently collapse and decay, are a common biological cause of deteriorating water quality. Water-supply problems associated with algal growth are discussed in the Australian context by Smalls (1980). The annual maximum chlorophyll-a concentration gives an indication of the worst that can be expected, or the goal to be sought if water quality is to improve. However, the breadth of the 95% confidence limits of such regressions causes concern. If the managerially desirable step of applying 95% prediction limits is taken, the range of chlorophyll-a concentration predicted for a given concentration of phosphorus is disturbingly broad (OECD 1982). This is demonstrated for Lake Burragorang and other regressions quoted here, by arithmetic ranges of predicted chlorophyll-a concentration (Table 7.3). For example, at site 3D an annual mean total phosphorus concentration of 20 mg m^{-3} predicts a maximum of 9 - 96 mg m^{-3}. This range may be reduced, if it is feasible to accept lesser degrees of predictive confidence. The OECD (1982) reported 80% prediction limits, and Dillon and Rigler (1974) reported 50% prediction limits, commenting that "...decisions about lake management often must be made when the chances of error are greater than 5%". In certain circumstances, it may be possible to take a "one-tailed" approach to the prediction limits; that is, to assert that underestimating the annual maximum concentration of chlorophyll-a is a more serious error than overestimating it.

FIGURE 7.3

Schematic representation of changes in regression slope resulting from changes in choice of regression sub-variables for Lake Burragorang and South African reservoirs [data from Walmsley and Butty (1980)]. Direction and magnitude of slope change is indicated by arrows and numbers (South African values in parentheses).



$y_{max} : \bar{x}$
0.54
(0.13)
$y_{max} : x_{max}$
0.37
(0.32)
0.17
(−0.19)
0.28
(0.29)
$\bar{y} : \bar{x}$
0.45
(0.10)
$\bar{y} : x_{max}$
Ln (chlorophyll concn)
Ln (total phosphorus concn)

In this case, there is a 1 in 10 chance of the annual maximum chlorophyll-a concentration exceeding the upper 80% prediction limit, and a 1 in 4 chance of exceeding the upper 50% prediction limit.

It is evident that the regressions, presented here, for Lake Burragorang are of limited use in forecasting events in the lake, as prediction of a chlorophyll maximum after the event (i.e. by the time an annual mean total phosphorus concentration is available) is of little benefit. It is, however, possible to gain some predictive insight. For example, the regression for the data from site 3D suggests a relatively great sensitivity to phophorus enrichment, in that the predicted annual maximum chlorophyll-a concentration (for annual mean total phosphorus concentration of 20 mg m^{-3}) is high compared to that predicted from the general OECD (1982) study or for South African reservoirs by equation (6) (Table 7.1). This may be of direct significance to future management of Lake Burragorang. Further, use of annual mean total phosphorus concentration, as the predictor variable (independent), lends itself to integration with phosphorus-loading models. These predict total phosphorus concentration in the lake on the basis of phosphorus influx, lake morphometry, water-retention time, and phosphorus retention by the lake. Such a combination provides a statistical model with which to assess the effect of projected nutrient loads or to formulate protective loading criteria for the lake. In the absence of detailed information on phosphorus loading for Lake Burragorang, and because some of the assumptions of such models are not met in this lake, the regressions presented here represent an initial step toward more detailed modelling of the chlorophyll-phosphorus relationship in a lake that retains a relatively undeveloped catchment.

In a broader sense, the results of this study support the informed application of Northern Hemisphere chlorophyll-phosphorus models to a cross-section of natural and artificial lakes in the Southern Hemisphere, although we note that a significant number of water bodies in South Africa and New Zealand have TN:TP ratios indicating potential nitrogen limitation. Approximately 24% (n =17) of South African reservoirs studied by Walmsley and Butty (1980) had TN:TP ratios <10, and White (1983) reported 37% (n = 27) of natural lakes in New Zealand with similarly low ratios. Bowles (1982) suggested that low ratios of nitrogen to phosphorus concentration indicated probable nitrogen limitation in some Victorian lakes. However, her use of the nitrate to total phosphorus concentration ratio would underestimate either of the commonly reported TN:TP or inorganic nitrogen to inorganic phosphorus concentration ratios. Data for dissolved phosphorus concentration, from Powling (1980), enable calculation of the nitrate to dissolved phosphorus concentration ratio, which should more closely approximate the inorganic nitrogen to inorganic phosphorus ratio. For Lake Eppalock and Lake Tarago, these are c. 50% higher than those reported by Bowles (1982) and fall within the range suggested by Forsberg et al (1978) to be nitrogen and/or phosphorus limited.

The informed use of Northern Hemisphere models in Australia and South Africa, at least, will often require consideration of the role of non-algal turbidity in modifying the chlorophyll-phosphorus relationship. Research into this effect will be of particular value in the management of eutrophication in the Southern Hemisphere.

SEASONAL CHANGES IN CHLOROPHYLL AT SITE 3D

The preceeding section establishes the likely significance of phosphorus limitation in Lake Burragorang on the basis of annual mean total phosphorus, and either the annual mean or annual maximum chloropyll-a concentration. Regression of annual maximum chlorophyll-a on matched total phosphorus concentration (single samples taken at the same time as those for chlorophyll-a determination) also shows a reasonably close relationship between the two variables (Note; in 8 of the 33 data pairs no exactly matched data for total phosphorus was available, so the second greatest chlorophyll-a record for the year and the matching total phosphorus determination have been used. For this reason, the subscript "max" is not used for the chlorophyll-a concentration). The regression explains 65% of the variation in chlorophyll-a concentration, and gives the following equation:-

$$\ln C = 1.006 \ln P_{matched} - 0.759 \quad (F = 59, p = 1.2\ 10^{-8})$$
$$\text{SE} \quad 0.131 \qquad\qquad 0.400$$

(Note: Here, F refers to the F statistic, and not to the correction factor reported in Tables 7.1 and 7.2.)

Monthly means of chlorophyll-a concentration, for the period from 1970 - 1980, are plotted in Fig. 7.4. The diagram is scaled (X-axis) as nearly as possible to that of the isopleth diagrams and the monthly total inflow minus evaporation plot (Chapter 3, Fig. 3.6).

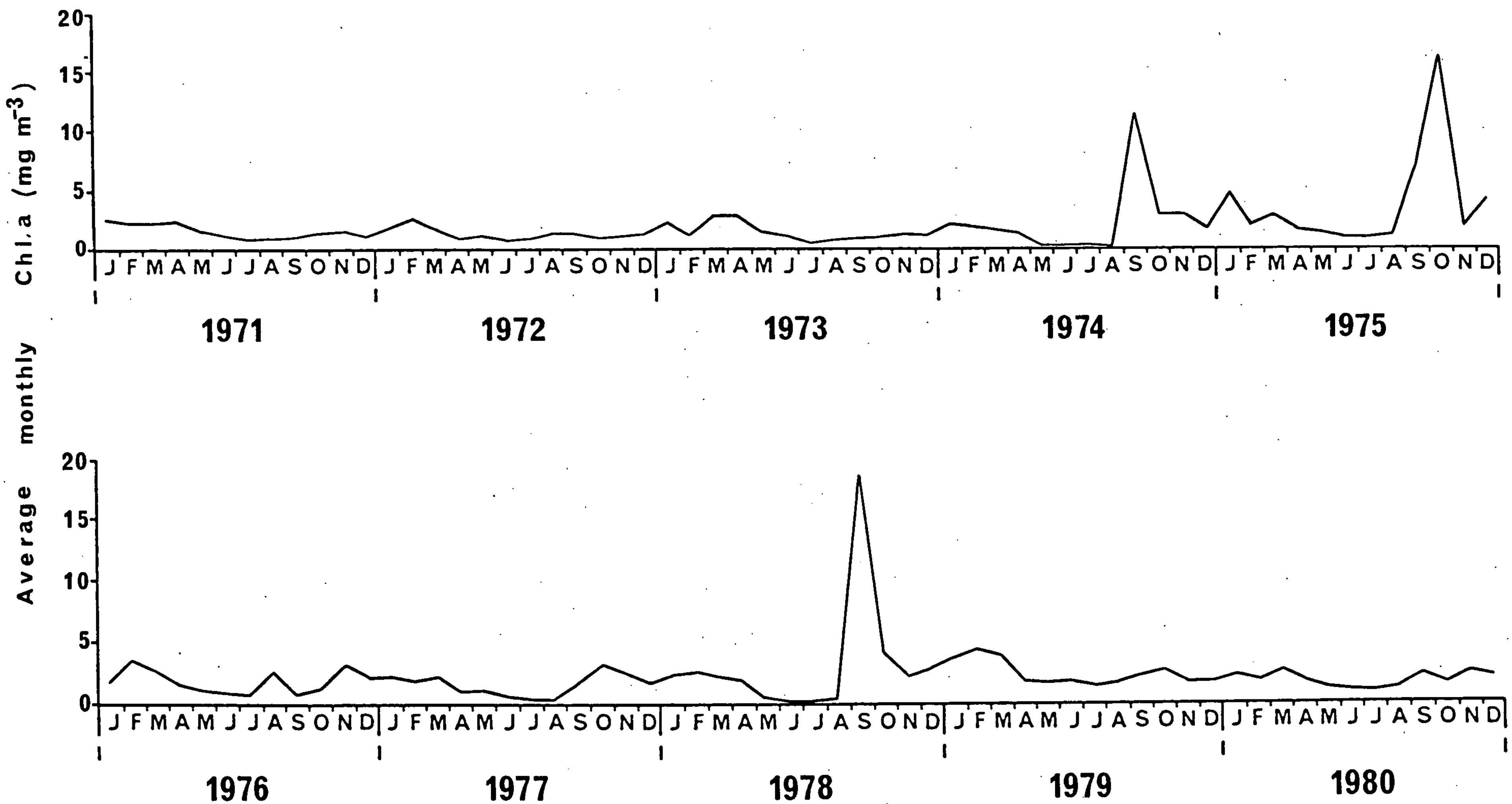
Fig. 7.4 shows the generally low levels of chlorophyll-a that prevail at site 3D. Concentrations greater than 5 mg m^{-3} are rare, which may be compared to the goemetric mean of annual euphotic chlorophyll-a concentration (8.4 mg m^{-3}) reported by the OECD (1982) for 96 lakes and reservoirs (mainly European and North American). The annual mean of chlorophyll-a at site 3D is 2.2 mg m^{-3} (n = 11; standard deviation = 0.76; range = 1.5 - 3.9: geometric mean = 2.3).

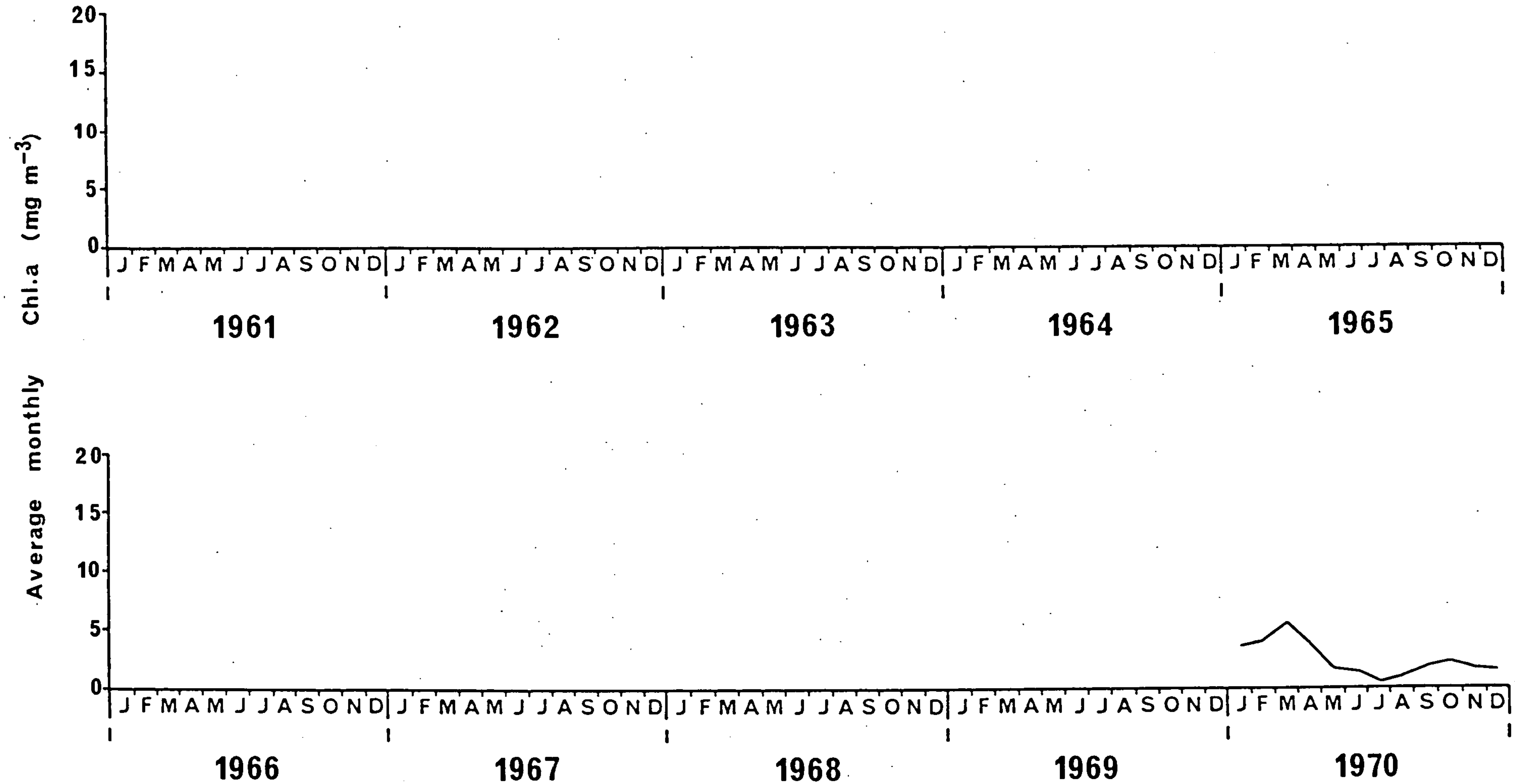
It was stated earlier, that the annual maximum concentration of chlorophyll-a could occur in any of about 10 months. This indicates a lack of clear seasonality in phytoplankton growth in this lake. Fig. 7.4 shows this poorly defined seasonality of biomass at site 3D, on the scale of monthly means of chlorophyll-a concentration. The most regular feature of the seasonal cycle is a depression of the chlorophyll-a concentration in winter, with the minimum concentration (0 - 4.5 m) usually occurring in July, when

FIGURE 7.4

Monthly averages of chlorophyll-a (mg m^{-3}) for the period 1970 - 1980 from euphotic zone (0 - 4.5 m) samples at site 3D. The time scale is as close as possible to that of the isopleth diagrams.

Three peaks of chlorophyll-a concentration have occurred in the study period, all occurred in spring, following significant inflows. All were dominated by the green colonial flagellates of the genus Volvox .


Average monthly Chl.a (mg m^-3)
20
15
10
5
0
J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D
1971
1972
1973
1974
1975
20
15
10
5
0
J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D
1976
1977
1978
1979
1980



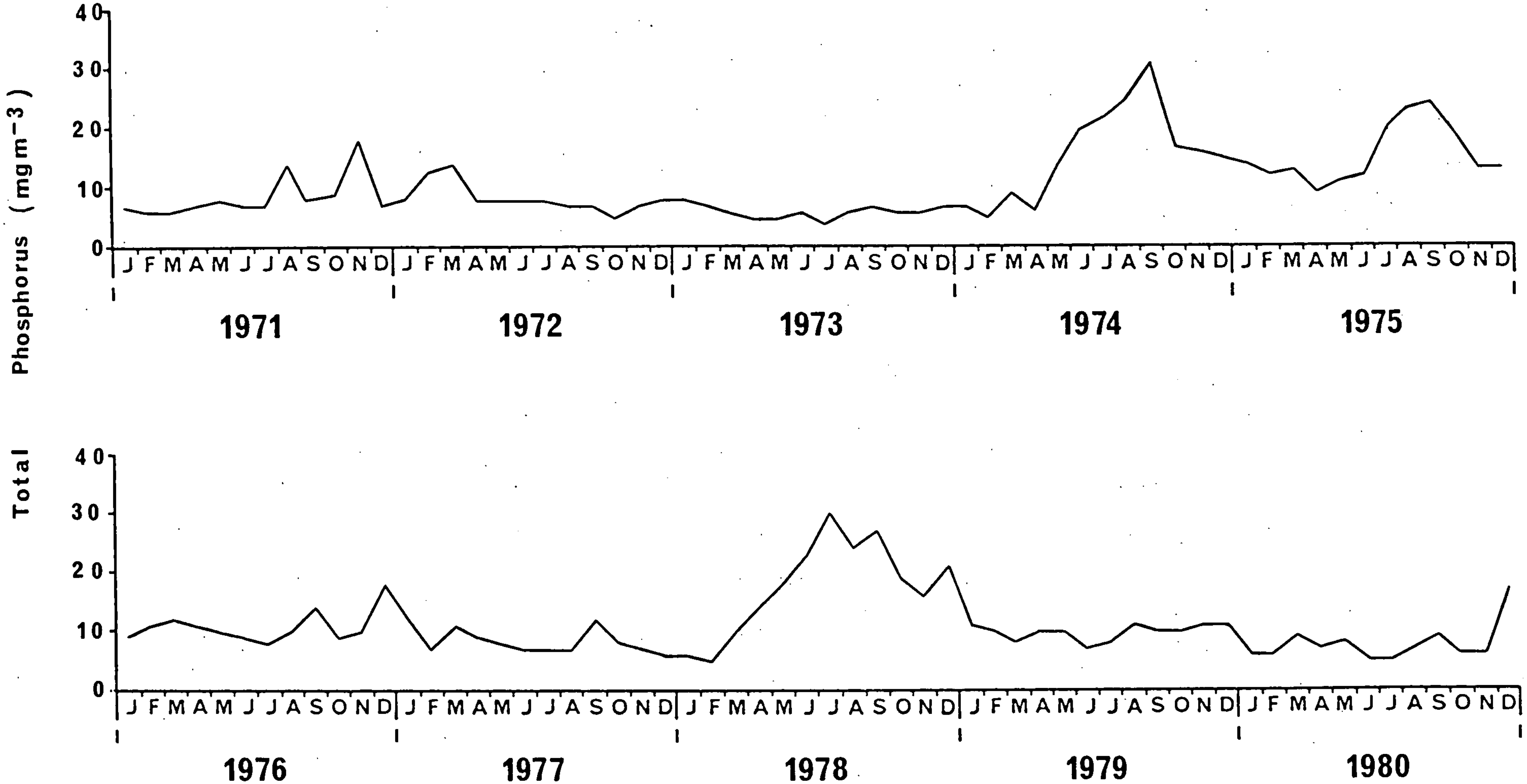
Average monthly Chl.a (mg m^{-3})
20
15
10
5
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

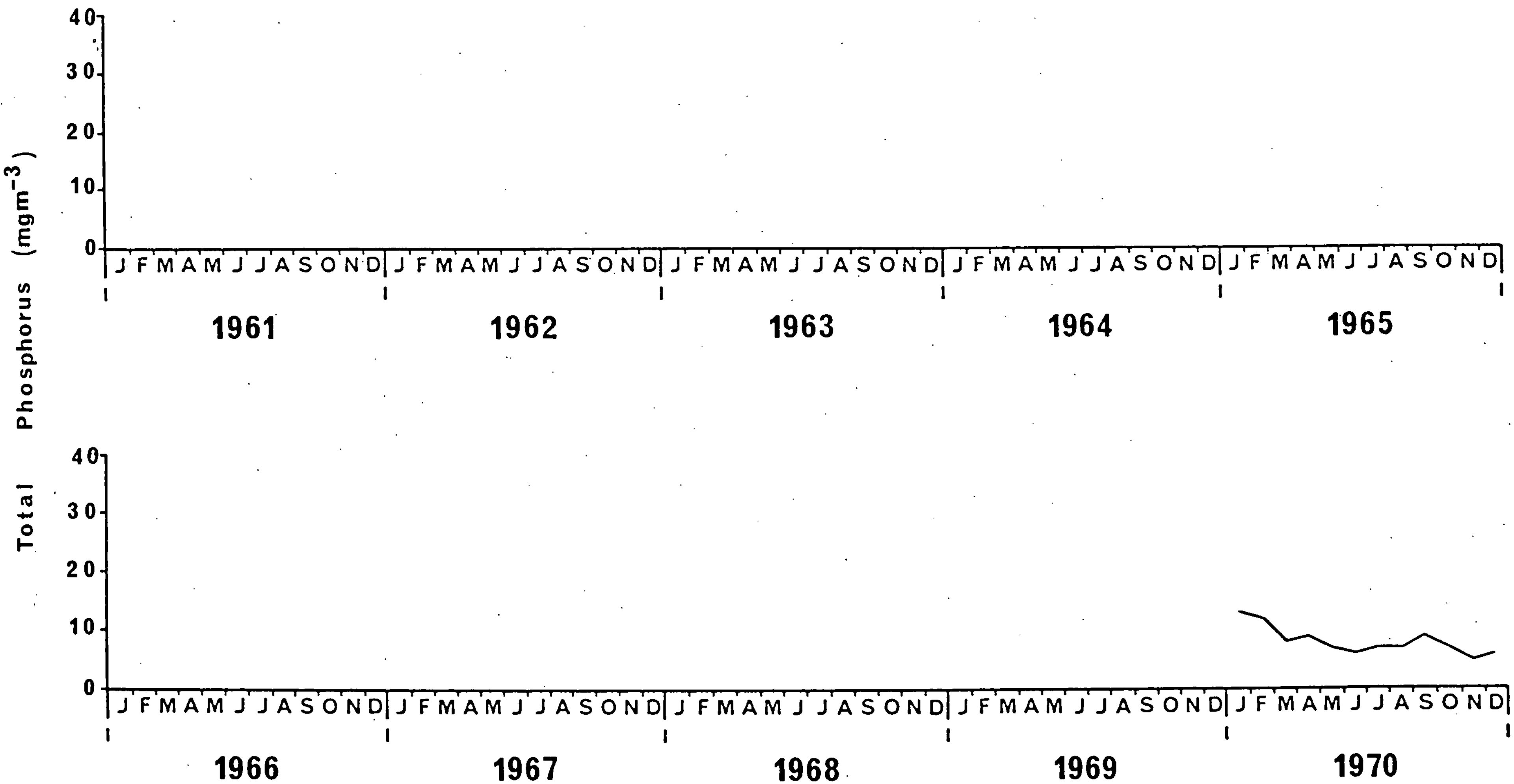
the mean concentration = 0.8 mg m^{-3} (n = 11; standard deviation = 0.37; range = 0.2 - 1.5). Direct comparison of Fig. 7.4 and the monthly inflow totals (minus evaporation; Chapter 3, Fig. 3.6) shows that very low winter concentrations of chlorophyll-a in 1974 and 1978 are more or less coincident with periods of high inflow. This probably indicates a loss of the algal population over the dam crest as a result of inflow, but some effect of turbidity is also likely (compare Fig. 7.4 with the turbidity isopleths Chapter 6, Fig. 6.1). Increasing turbidity, combined with the deepening mixed zone would cause the underwater light climate to deteriorate rapidly by decreasing the average residence time of phytoplankton in the euphotic zone (see Talling 1971). The potentially limiting effect of light climate on phytoplankton growth has been demonstrated by Ferris (1977; reported in Ganf 1980) for Mount Bold Reservoir (South Australia).

Monthly average chlorophyll-a concentration exceeded 10 mg m^{-3} on only 3 occasions between 1970 and 1980 (Fig. 7.4). Each of these blooms, which occurred in September/October (Spring) of 1974, 1975, and 1978, was dominated by motile, green, colonial flagellates of the genus Volvox. Fig. 7.5 shows monthly means of total phosphorus (0 - 4.5 m; mg m^{-3}), and it is evident that the algal blooms followed periods when total phosphorus concentration exceeded about 20 mg m^{-3}. It is also apparent from Figs 7.4, 7.5, 3.6 (total inflow, Chapter 3), and 6.1 (turbidity isopleths, Chapter 6) that these three algal blooms at site 3D followed circulation periods when turbid water from pre-circulation inflows was mixed into the euphotic zone. The fact that these blooms developed under relatively adverse light conditions, supports the contention that after the reformation of the thermocline (i.e. when a relatively shallow mixed zone is re-established), it is the availability of nutrients, particularly phosphorus, which is critical in limiting algal growth at site 3D. Inflows, it seems, have antagonistic effects on the conditions which determine phytoplankton growth. Nutrients are provided, but

FIGURE 7.5

Monthly averages of total phosphorus concentration (mg m^{-3}) from euphotic zone (0 - 4.5 m) samples taken at site 3D in the period 1970 - 1980. The time scale is as nearly equivalent to that of the isopleth diagrams as possible. The major peaks of phosphorus concentration coincide with the three peaks of monthly average chlorophyll-a (see Fig. 7.4).


Total Phosphorus (mgm⁻³)
40
30
20
10
0
J F M A M J J A S O N D
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980


Total Phosphorus (mgm⁻³)
40
30
20
10
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

at the same time light is restricted.

INFLOW AND PHOSPHORUS CONCENTRATION

In view of the generally close relationship between phosphorus concentration and phytoplankton biomass (as chlorophyll-a), the prediction of phosphorus concentration at site 3D is a logical step. Comparison of Fig. 7.5 with the monthly total inflow minus evaporation (Chapter 3, Fig. 3.6) and the isopleths of turbidity (Chapter 6, Fig. 6.1) indicates that turbidity and inflow volume may explain a significant percentage of the variation in total phosphorus concentration at site 3D. Bowen and Smalls (1980) report a significant correlation between annual total inflow (minus evaporation) and the mean annual total phosphorus concentration in Lake Burragorang (1972 - 1974). It is well established that phosphorus is frequently associated with suspended particulate matter carried in streams (Barlow and Glase 1982; Pierrou 1979), particularly with the clay fraction Golterman (1973). Oades (1982) concludes that during storm flows "... most of the phosphate in the water is associated with fine colloids and is not in true solution.", indicating that even that fraction of the total phosphorus generally regarded as dissolved, may in fact be associated with fine particles.

These findings support the likelihood that inflow and suspended load (roughly indicated by turbidity) will be important in determining the total phosphorus concentration in Lake Burragorang. Another possible source of phosphorus is from agriculture in the catchment of the lake. According to Cook and Williams (1973) the greatest loss of agriculturally applied phosphorus is by erosion of soil with bound phosphorus, again, pointing to the potential significance of suspended particulate matter.

Regression Analysis

In the following analysis, 10 years of data for annual mean total phosphorus at site 3D is regressed on up to three independent variates. These are:-

1. Total annual inflow minus evaporation (x $10^6 m^3$).
2. The maximum surface turbidity recorded at site 3D during the year.
3. Tonnage of superphosphate applied to farm land in the shires which include the natural catchment of the lake. Note that these shires extend beyond the catchment also, so that only a percentage of the reported tonnages of fertilizer were actually used within the catchment. I have assumed that this was a constant proportion of the total. This variate is also lagged by one year as part of the analysis.

Superphosphate and Total Phosphorus Concentration:-

No significant linear relationship is found between superphosphate application in the catchment and the mean annual total phosphorus (mg m^{-3}) in the euphotic zone at 3D, whether the current or previous year's application of superphosphate is considered.

It was thought that some effect of this variate might be found, as a result of the very significant changes in fertilizer application that occurred during the course of the study period. This change was brought about by a political decision to abolish the subsidy ("Super Bounty") paid to farmers to encourage the use of phosphate fertilizer. The decision came into effect in December 1974 (Australian Fertilizers Ltd., Sydney; pers. comm.) and in the year prior to that record applications of superphosphate occurred (c. 138% of mean levels for 1963 - 1972). Then, in the next two years, levels of superphosphate application fell to those of the mid-1950's (c. 38% of mean levels for the period 1963 - 1972). When the subsidy was re-introduced, in January 1976 (Australian Fertilizers Ltd., Sydney; pers. comm.), the tonnage of superphosphate recovered to the more recent levels, over a period of about 3 years (Data from the N.S.W. Bureau of Statistics).

It is evident, however, that no simple relationship exists between the application of fertilizer in the area including the physical catchment of Lake Burragorang and the concentration of total phosphorus in the lake (at site 3D). This does not imply that changes in agricultural practice, of the magnitude reported above, do not affect Lake Burragorang, only that any effect of catchment events on the lake is subject to the intermediary effect of inflow.

Inflow, Turbidity and Total Phosphorus Concentration:-

Extending Bowen and Smalls' (1980) analysis to include 11 years (1970 -1980) provides the following relationship, which explains 70% of the variation in annual mean total phosphorus concentration at site 3D (untransformed data conforms with the requirements of the linear regression model):-

$$\ln \bar{P}_a = 0.00259 \ln \text{Vol} - 7.137 \quad (F = 20.6,\ p = 0.0014)$$
$$\text{SE} \quad 0.00057 \qquad 1.043$$

Fig. 7.6 shows this regression with 95% confidence and prediction limits. Despite the fact that the regression is significant at the 1% level ($p < 0.01$), the prediction limits are quite broad.

The maximum record of surface turbidity explains 72% of the variation in annual mean total phosphorus concentration at site 3D. The regression equation for 10 years data (1970 - 1979) is:-

$$\ln \bar{P}_a = 0.473 \ln \text{Turb} - 7.719 \quad (F = 22.9,\ p = 0.0099)$$
$$\text{SE} \quad 0.099 \qquad 0.907$$

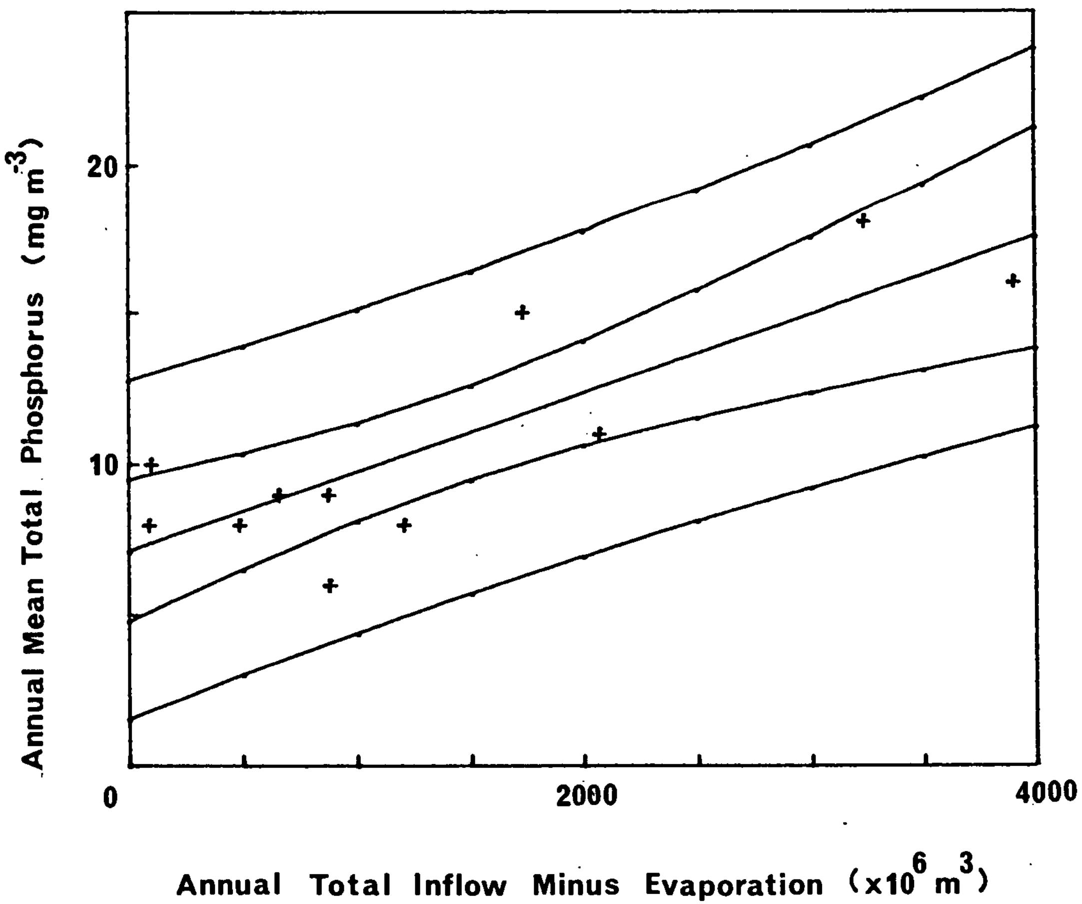
This relationship is shown in Fig. 7.7. The distribution of points along the regression line is not particularly good because most of the points fall at the lower end of the turbidity scale. The regression is, therefore, primarily

FIGURE 7.6

Linear regression of annual mean total phosphorus (mg m^{-3}) on annual total inflow minus evaporation (10^6 m^3) for data from 1970 - 1980. The variates are not transformed. The 95% confidence (inner line) and predictive (outer line) limits are marked on the plot. The regression explains 70% of the variation in annual mean total phosphorus concentration at site 3D (n= 11). Despite the fact that the regression is significant at the 1% level ($p < 0.01$), the prediction limits are quite broad.

$$\ln P_a = 0.00259 \ln Vol - 7.137 \quad (F = 20.6, p = 0.0014)$$

SE 0.00057 1.043

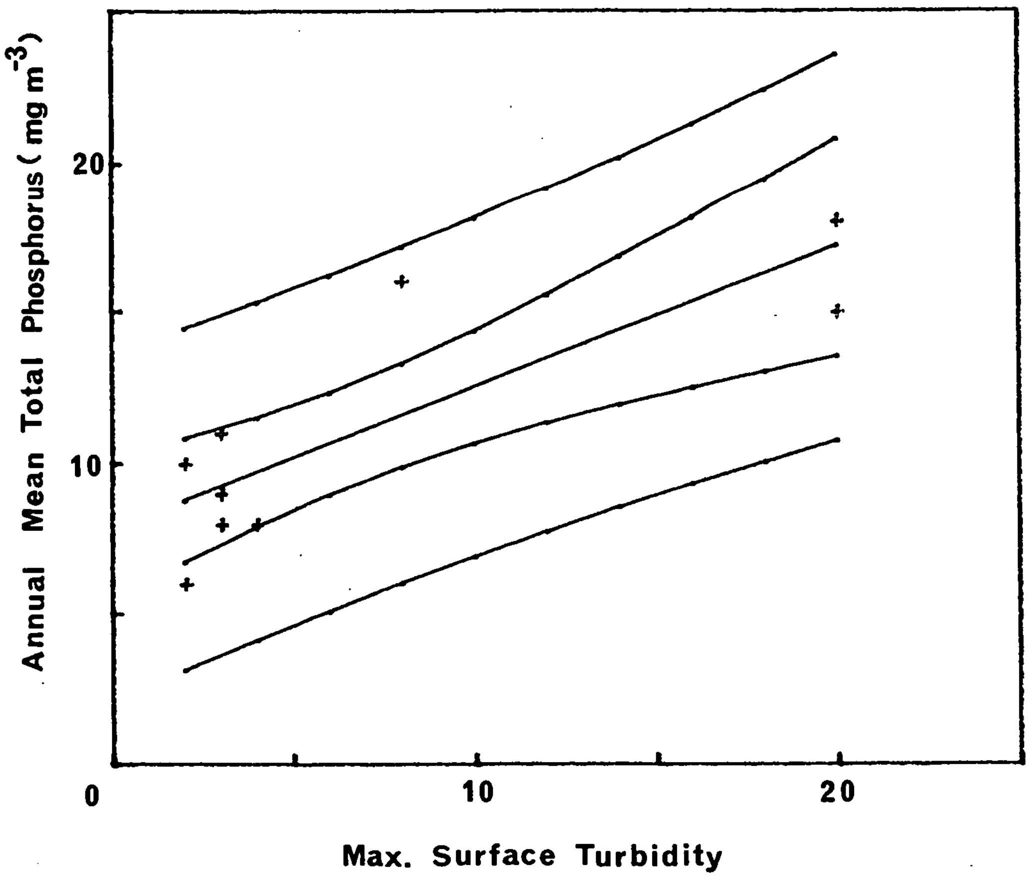


Annual Mean Total Phosphorus (mg m-3)
20
10
0
2000
4000
Annual Total Inflow Minus Evaporation (x10 6 m 3)

FIGURE 7.7

Regression of annual mean total phosphorus (mg m^{-3}) on maximum surface turbidity from each of the years 1971 - 1979. Data is not transformed. It is evident that the prediction limits are quite broad. The grouping of most points at the lower end of the turbidity scale means that the regression is primarily determined by very few data points, and this is not particularly satisfactory. The maximum record of surface turbidity explains 72% of the variation in annual mean total phosphorus concentration at site 3D (n = 10).

$\ln P_a =$	$0.473 \ln Turb$	$- 7.719$	($F = 22.9$, $p = 0.0099$)
SE	0.099	0.907	


Annual Mean Total Phosphorus (mg m^{-3})
20
10
0
10
20
Max. Surface Turbidity

determined by very few data points. It should be mentioned that the turbidity method (Hellige) has been discontinued, and the relationship shown in Fig. 7.7 cannot be used with the new readings. The regression on total inflow (Fig. 7.6) is, therefore, the preferred regression.

There are insufficient data to calculate a phosphorus budget for Lake Burragorang because no data exist for the concentration of phosphorus in the Wollondilly River (for the present study period), although this is arguably the most significant source of phosphorus to lake Burragorang in terms of total load. Monitoring of phosphorus concentration has so far been confined to the Cox River and Kedumba Creek because of the point source of nutrients (Katoomba Sewage Treatment Works) which discharges into Kedumba Creek.

A problem of practical significance to the managers of Lake Burragorang is to distinguish between any eutrophicating trend in the lake and the year to year variation driven by advective processes. The apparently straightforward relationship between inflow volume and total phosphorus, linked to the phosphorus-chloropyll model, provides an approximate relationship with which to predict expected chlorophyll-a concentrations for a given inflow volume. This could facilitate the detection of any trend towards enrichment of the lake, which may otherwise be obscured by the advectively driven fluctuations in nutrient concentration.

PART IV

Chapter 8

INTRODUCTION

In the final chapter of the thesis some consideration is given to whether the behavioural variation described for Lake Burragorang, on the basis of a twenty year monitoring period, has any directional component caused by the evolution of the lake since it filled, or by eutrophication. It is important to assess whether the variable behaviour of this lake is essentially "natural" and generated mainly by the unpredictable climatic factors that govern the flow of many rivers on Australia's east coast (see Hart 1974; Williams 1982), or whether the lake has varied with time and in response to man-made disturbance. If this question can be answered, it further defines the context within which to view the variation that has been described for this twenty year study period.

The implications of the study for the management of Lake Burragorang are also considered briefly, and some suggestion for experimental work that could enhance the value of the present data body is described.

CHAPTER 8

CONCLUDING DISCUSSION

LONG-TERM BEHAVIOURAL VARIATION

Advection and Variation

Variation, on the scale of years, is a central theme of the present work. Lake Burragorang clearly exhibits considerable variation in many important aspects of its physical, chemical, and biological cycles, with time. The lake's thermal pattern, though essentially warm monomictic, is interrupted by years in which the winter circulation is incomplete. The vertical distribution of heat varies between years (most noticeably during the cooling phase), from a situation in which the thermocline (averaged for each month) is maintained within about 20 m of the lake surface throughout the summer and most of the autumn, only deepening in the final month or two prior to the winter period of maximum vertical circulation, to behaviour characterised by a steady deepening of the thermocline from mid-summer to autumn, such that it may occupy a position some 20 m deeper (by April - May) than in the previous sequence. Related to this, is the finding that the annual heat budget has an unusually high coefficient of variation, compared to others of the world's lakes.

The cycle of oxygen stratification is also variable, as indicated by the range of volumetric hypolimnetic oxygen depletion rates (VHDR), recorded in the study period. Corrected to a uniform temperature (4°C), these range from 0.11 - 0.89 mg l^{-1} $month^{-1}$, equivalent to 39% of the range of Vollenweider and Janus' (1984) data from 21 European and North American lakes (without temperature correction, the range of Lake Burragorang's VHDR is equivalent to 60% of the uncorrected range reported by Vollenweider and Janus 1984).

The importance of inflow and outflow, in generating much of this behavioural variation, has been established by a comparison of selected WET and DRY years. The analysis identified a number of aspects of the stratification cycle (temperature and dissolved oxygen) that may be specifically attributed to the effects of inflow and outflow, although this comparison, based on approximately uniformly wet or dry years, could not account for the full range of variation that might be caused by combinations of wet and dry year characteristics within any one year, and did not include data for the years in which the largest floods occurred.

Of the points mentioned above, the interruption of the monomictic cycle results from a combination of the stabilising effect of cold underflows, which maintain a thermal gradient and measurably increase the Schmidt stability during the winter, and the protection of the deeper layers by the withdrawal current associated with the mid-depth hydro-electric offtake. This current probably acts as a barrier to vertical convective mixing below the offtake depth, as suggested for Lake Powell (U.S.A.) by Johnson and Merritt (1979), and shown for Dartmouth Reservoir (Victoria) by Welsh (1984). The hydro-electric offtake (combined with metalimnetic inflow) actively promotes the downward movement of heat in the overlying water column, causing the steady deepening trend of the thermocline found in the summer and autumn of wet years, and increasing heat storage in the lake. The subsequent promotion of heat loss, throughout the water column, when inflows begin to underflow the lake water, results in a similar winter minimum heat content for both wet and dry years, so that the effect of advection is to increase the annual heat budget, and therefore the year to year variation of this heat storage term.

Advection contributes to a greater rate of hypolimnetic oxygen depletion, through the introduction of turbid water into the lake by inflows, and, in the case of outflows, by drawing less oxygenated water, from shallower upstream sites, into the water column at site 3D. In the comparison of WET and DRY year groups, the WET year VHDR (January - May) was about twice that of the DRY years.

A substantial part of the variation in algal biomass (as chlorophyll-a, at site 3D) can be attributed to the effects of inflows, in that the few algal blooms, recorded at site 3D since 1970 when chlorophyll measurements began, followed inflows which brought particulate associated nutrients (especially phosphorus) into the lake. These blooms occurred in spring, after the winter circulation distributed the influent turbidity into the upper water column, and despite the decrease in light penetration that resulted from this. The dominant role of advective supply of nutrients in determining algal biomass has also been observed for Lake Powell (U.S.A.) by Gloss et al 1980.

Alternative Sources of Variation

Having identified the important contribution of inflow and outflow to the range of behaviour recorded for Lake Burragorang, some consideration must be given to the possibility that some of this variation is directional, representing trends associated with progressive eutrophication of the lake, or with the filling and equilibration stages reported for many newly filled impoundments (see Balon 1974; Baxter 1977).

Trophic upsurge:-

This process, which accompanies the flooding of terrestrial soil and plant material with the subsequent death and decay of the terrestrial organisms, and the leaching of material from the flooded soil, generally involves an increase in the concentration of total dissolved solids and nutrients (Zhadin and Gerd 1961; Balon 1974), and is associated with increased oxygen consumption in the new hypolimnion (Steane and Tyler 1982, Tyler and Buckney 1974, Welsh 1984). The accompanying biological response usually includes temporarily increased production of algae (Baxter 1977), and fish (Petr 1975; Zhadin and Gerd 1961). The timing of these changes, and the period required for each factor (chemistry, primary and secondary production) to achieve some equilibrium, differs between factors, and according to various circumstances specific to each lake (ie lake depth, water retention time and temperature, and the length of the filling period).

According to the schematic representation of Balon (1974), chemical stabilisation, occurs more quickly than that of biological production, taking about 6 years in Lake Kariba (see also Coche 1968). Baxter (1977) reviewing earlier work, reports a period from about 1 - 6 years for the chemical equilibration of water with newly flooded soil, and in laboratory experiments, Gunnison et al (1980) found that three simulated cyles of oxic/anoxic conditions were sufficient to approximately halve the release rates of some soluble nutrients (ammonium-nitrogen, and orthophosphate-phosphorus) from flooded soil.

The surge of biological production took about 10 years to decline in Lake Kariba (Balon 1974). Petr (1975) lists Lakes Volta, Kariba, and Kainji, as apparently stabilising after 4 - 15 years, based on commercial fish catches.

A preliminary consideration is to set the study period in the context of the history of the lake's formation. A small lake (c. 15 m deep) has existed in the narrow Warragamba Gorge since 1940. Although Warragamba Dam was not closed until 1960, the gradual formation of Lake Burragorang probably dates back to 1957 (when the water diversion tunnel was closed), or earlier according to Jolly (1966a), who refers to water storage beginning in 1956. Mayrhoffer (pers. comm. 1982) suggests that the lake depth may have been independent of the small coffer dam, even earlier (c. 1955). Therefore, it seems likely that a period of at least 4 - 5 years, of the filling phase, is not represented in the data given in this thesis. Nevertheless, the maps (see Chapter 1, Fig. 1.1) showing the extent of the lake in February 1959 and November 1961, indicate that a substantial area of land was flooded after 1959, and so that ~~therefore,~~ the present data set should include some evidence of any post-impoundment increase in dissolved solids, or tendency towards greater oxygen depletion in the hypolimnion.

In fact, there are elements of the lake's chemical behaviour, following the final filling of the lake, that could be interpreted as evidence of the predicted sequence, particularly from a short-term (say 3 - 4 years) viewpoint. In the period from 1962 - 1965, there is progressively less severe hypolimnetic de-oxygenation, as evidenced by declining volumetric hypolimnetic oxygen depletion rate (VHDR), and shorter periods of near anoxia at site 3D, until in 1965 the hypolimnion remained more than 20% saturated with oxygen for the whole year. At the same time, there was a general improvement in water clarity, with a reduction in total iron concentration (and presumably in phosphorus). This could be explained, at least in part, as the period of chemical decrease, following the initial trophic upsurge, which accompanied the gradual filling of the lake in the previous 3 - 5 years.

The changes which took place between 1962 and 1965 may, however, be simply explained as the effects of the November 1961 flood, which filled the lake for the first time, and was followed by comparatively wet years from 1962 to 1964. The flood of November 1961 is unique in the context of this study, both for its timing, and for its size (monthly total inflow volume) and intensity. The total volume (minus evaporation) exceeded 1600 10^6 m^3 for November 1961, and is among only 7 monthly inflow totals to exceed half this figure, in the period from August 1961 - December 1980. Most of these large inflows occurred in autumn or winter, with only one other November flood, that of November 1969, which totalled c. 800 10^6 m^3. The intensity of the November 1961 flood, can be judged from the fact that the water level rose 2.79 m above the dam crest at peak flow, subsequently, the flood in June 1964 (the second largest in volume, for the study period) peaked at 2.19 m above the dam crest, while later flows have risen less than 1.5 m above the dam crest. The following passage, from the 1965 Warragamba Catchment Area Detailed Erosion Survey (Soil Conservation Service 1965), highlights the effect of the 1961 flood from the viewpoint of soil erosion in part of the

Wollondilly River catchment:-

"Severe flood rains which fell in the Twin Peaks - Mt. Yerranderie area in November 1961 (12 inches in 24 hours) caused tremendous landslips and scouring in the drainage lines below these mesa-like plateaux. The rock debris and water gouged large channels up to two chains wide and fifteen feet deep along normally small intermittent water courses. This silt and debris passed on to the streams further down and possibly into the dam. A feature of this erosion, however, is that it was completely unpreventable. It was not the result of wash on a burnt area but merely the combination of topography and weather."

The sediment carried by this storm obviously did reach the lake, causing water quality changes that have not been repeated in later years. The inflow reached site 3D as an underflow, increasing the hypolimnetic temperature to > 15°C, and the turbidity to 700 Hellige units. In subsequent years, hypolimnetic temperature has remained less than c. 13.5°C, and turbidity less than 300 Hellige units. The combination of high turbidity and a relatively warm hypolimnion, probably accounts for the rapid depletion of oxygen early in 1962, leading to the only period of anoxia at site 3D. The progressive decrease in VHDR, and the general improvement of water quality may represent the combined aftermath of the flood, and the changes that accompanied the dryer conditions of late 1964 and in 1965.

Overall, although the circumstances that surrounded the final filling of Lake Burragorang, and conditioned the physical and chemical behaviour in the early years of this study, were unique in the context of the available data, there is no compelling reason to ascribe these events to post-impoundment enrichment of the lake, as they can be explained in terms of the known effects of advection. This conclusion is supported by Jolly (1966a), who examined net samples of phytoplankton and zooplankton from Lake Burragorang, in the period September 1959 to 1964, and observed "... that there was no burst of productivity during the first impoundment years.". In drawing this conclusion, there is also the tacit assertion, that the physical and chemical events that accompanied the 1961 flood, may be repeated in the lake if the same combination of unusual circumstances occurs again. It should also be noted, that I do not discount the possibility of some chemical or biological effect associated with the filling phase in Lake Burragorang, but I consider that any such effects were small in comparison to the effects of the large inflow that filled the lake. Neither this study, nor that of Jolly (1966a) covered the complete period of the lake's formation.

In looking for reasons why Lake Burragorang did not show an obvious trophic upsurge, a significant fact relates to the site preparation prior to, and during, the filling stages. Mouchet (1984) cites various examples of the effectiveness of site preparation in reducing water quality problems in newly filled impoundments. The region to be flooded was cut and burned in advance of the rising water (see Historic Perspective, Chapter 1), and although there was no topsoil stripping, a factor that has been suggested by some (Gunnison et al 1980) to be of greater significance than de-vegetation, much of the region is characterised by phosphorus-poor soils, derived from the predominantly sandstone rocks of the catchment (see Geology, Chapter 1; Beadle 1954, 1962; Soil Conservation Service 1962, 1965), and the flooding of such soil may not have contributed much to the nutrient load in the lake. Beadle (1962) examined the phosphate content of the Hawkesbury sandstones, Wianamatta and Narrabeen shales, and the soils derived from these rocks, and found relatively little variation of phosphate content among the surface soils, in contrast to the parent material, and that the phosphate content of the soils was much lower than the mean content of the parent rocks, a fact he attributed to the large amount of phosphate held in the vegetation, mostly above ground. Therefore, the removal of the vegetation (by clearing and burning) may have been quite effective in reducing the release of this nutrient following flooding of the lake basin. Nevertheless, only part of the cut timber was actually removed from the Lake Burragorang basin, the rest was burnt in situ, with the cleared understorey vegetation. Mouchet (1984) points out that this practice simply ensures the rapid release of nutrients from the ash, into the rising lake water. However, the clearing process for Lake Burragorang lasted about 9 years, during which numerous small floods would have contributed to the passage of such nutrients, through the small lake. Certainly, there would not have been any single massive release of nutrients from this source.

It is evident from consideration of the nitrogen to phosphorus ratio, and the finding of a close log-linear relationship between chlorophyll and phosphorus in the lake, that phosphorus is the most likely limiting nutrient in the system today, and it is therefore reasonable to conclude that site preparation was effective in reducing any post-flooding nutrient increase in Lake Burragorang.

Another factor, specific to the present data set (ie from 1961 onwards), may have been the flushing effect of the November 1961 inflow, and the deposition of sediment in the reservoir, which could have prevented further release of material from the flooded soil and vegetation; this latter effect was also suggested by Jolly (1966a).

The preceeding discussion sets out the reasons for my conclusion that no period of trophic upsurge is apparent from the avaliable data, either chemical or biological, and that the unusual conditions of turbidity, and oxygen depletion (in relative terms), are adequately explained as the products of the largest and most intense flood recorded for the lake. That such an inflow should have filled the lake, is coincidental, but certainly complicates the interpretation of the early record of the lake's behaviour, and may have masked, or truncated a phase of trophic upsurge. The initial phase of the lake's history, may have been completely different (chemically and biologically) had the lake been filled by a lesser inflow.

Eutrophication:-

A voluminous literature describing the changes associated with eutrophication is available today, and in a specifically Australian context, Wood (1975), O'Loughlin and Cullen (1982), and AWRC (1983) are representative of the development in our understanding of eutrophication in the past decade. Arguably, one of the more important developments in the study and management of eutrophication in the world, has arisen from the limiting nutrient concept (Vallentyne 1974), and the development of simple nutrient concentration - biomass models, and their subsequent coupling to nutrient loading models (see Vollenweider 1968; Rast et al 1983). Phosphorus concentration, in particular, has been found to limit phytoplankton biomass in a fairly wide variety of lakes, in many places on earth (see Chapter 7), and phosphorus is also the most readily manipulated of the two nutrients (nitrogen and phosphorus) that are generally regarded as most likely to limit algal growth.

Considering the significance of phosphorus limitation at site 3D, in Lake Burragorang, probably the most important direct evidence for any trend towards eutrophication in Lake Burragorang (within the data considered here), are chlorophyll-a and phosphorus concentrations, which have been measured since 1970. For earlier years there is indirect evidence from trends in volumetric de-oxygenation, and determinations of nitrate concentration for the water drawn off to supply Sydney. Unfortunately, the data for nitrate concentration, is broken by a method change in 1974 (A. Robinson pers. comm.), which altered the sensitivity of the determination, increasing values by 5 - 8 times. I am unaware of any attempt to intercalibrate between the two methods, to determine if a simple relationship can be applied to the conversion of the results.

Of the data presented as part of this study, the hypolimnetic oxygen concentration and rates of depletion, show no trend that is apparent above the variation that can be attributed to naturally changing inflow (and therefore outflow) volumes for the lake (see Chapters 4 and 5). This is also true of the data for euphotic (0 - 4.5 m) chlorophyll and phosphorus concentrations, in the period from 1970 - 1980 (see chapter 7). Bowen and Smalls (1980) observed that the input of phosphorus to Lake Burragorang (site 3D) was greatly increased in response to stormwater flows, and although this could have been interpreted as a trend towards eutrophication, on the basis of the data collected to that time (1970 - 1977), subsequent data indicate that with a reduction of inflow (particularly in 1979 - 1980), the concentration of phosphorus declined to levels similar to those of 1970 (a relatively dry year).

There has been some discussion, of the effects of bushfires on Lake Burragorang (see Cullen 1983), in terms of increasing algal numbers after the 1968 fires (see Chapter 1), but, as phosphorus measurements commenced in 1970, and in view of the lack of any relationship between chlorophyll-a concentration and total algal cell counts (regression analysis by J. Ferris) for the period 1970 - 1980, I consider the data to be inconsistent with the main body of information presented in this study. The question of the effects of bushfires, is therefore unresolved, but there are reasons to assert that any influence they might have on Lake Burragorang may be lessening with time, as a result of more effective fire control measures. The importance of fire control, as a means of preventing erosion in the catchment, was stressed in the report of the Soil Conservation Service (1965).

From the preceeding discussion, I consider that the indirect evidence from hypolimnetic oxygen concentration in the years prior to 1970, and the direct measurement of algal biomass (as chlorophyll-a) and phosphorus concentration in the later years, indicate that Lake Burragorang was not subject to any detectable eutrophication within the study period. It must be noted, however, that the dominance of advectively supplied nutrients (at site

3D), and the unpredictable timing and magnitude of inflows, may make it difficult to detect such a trend.

Management:-

A source of some directional change, that I have only briefly alluded to, is that of management procedure for Warragamba Dam itself. I have mentioned some aspects of catchment managerial practice that may have affected the lake since its formation, specifically, management of the inner catchment, to prevent fire and direct agricultural practice, in order to reduce soil erosion. Also, there have been some less directed cultural factors mentioned, such as changing economic climate and problems with vermin, which have affected land use within the catchment. The most important aspect of the management of Warragamba Dam is that of the policy governing use of the HEPS offtake. The limnological significance of this outlet is detailed in Chapters 4 and 5. It is evident from a study of daily records of outflow from Warragamba Dam that the use of the HEPS has changed with time, becoming more responsive to the fluctuation of inflow volume. In the first four or five years of operation the HEPS was less strictly tied to fluctuations in the lake level than in later years. This can be seen by comparing Fig. 4.10 (HEPS offtake volume, Chapter 4) with Fig. 3.6 (total inflow minus evaporation, Chapter 3). The relative changes of inflow volume are similar for 1963 and 1974 (Fig. 3.6). Fig. 4.10 shows that the HEPS outflow volume is more nearly proportional to the inflow in 1974, than is the case for 1963. The point is simply that the early years in the life of Lake Burragorang were a period of adjustment in a managerial sense, as well as in terms of the lake's limnological behaviour. In view of the significant effect that management of the dam's outflow facilities has on the lake, especially at site 3D, this may be a factor of greater importance than its treatment here suggests, both in this lake and other impoundments.

Conclusion:-

Overall, I conclude that the physical, chemical, and available biological data for the twenty year study period (excluding data on phytoplankton cell numbers), is largely representative of the natural variation in Lake Burragorang's behaviour, and does not include any significant trend resulting from either the initial stages of the reservoir's life, or any progressive eutrophication of the lake. The considerable variation of Lake Burragorang's behaviour (indicated from data for site 3D) is driven primarily by non seasonal and annually variable inflow (and sub-surface outflow which is correlated with it). The present study, therefore, may be regarded as a baseline study of variability in a system presently reserved from much development within its catchment, but subject to perturbation by comparatively great climatic uncertainty, which is also a feature of other regions in Australia (Hart 1974; Williams 1982). The conclusion that this study is largely descriptive of natural variation in a comparatively variable system is important in relation to changes that have occurred, or may occur, in the use of Lake Burragorang.

First, the recent development of the Shoalhaven Scheme has linked three upstream impoundments with Lake Burragorang, so that it can receive water from outside of its natural catchment when its level falls to a pre-determined extent (Petrie and Smalls 1981). Bowen and Smalls (1980) indicate that the concentration of phosphorus in two of these upstream storages exceeds that recorded at site 3D by a factor of two to three. This linkage may, therefore, be expected to induce significant changes in Lake Burragorang, depending on the volume and timing of releases into the lake.

Second, there is increasing pressure in Australia for recreational access to lakes such as Lake Burragorang (Burton 1982), which is both a beautiful lake and situated close to a city of more than 3 million people. The degree of protection presently afforded Lake Burragorang and its inner catchment, although less than that for some other Australian water supply reservoirs, is remarkable in comparison with many water supply reservoirs in Britain and the United States (cf. Burton 1982). It seems likely that increasing recreational use of lakes such as Lake Burragorang will be inevitable in coming years.

The present study of Lake Burragorang provides an organised account of the lake and its capacity to vary over time, against which to measure changes that arise from linkage of the lake into a multi-reservoir system and potential increase in public access to the lake and its catchment.

However, most changes that presently occur in the catchment of the lake, with the exception of point sources of nutrient input like the treated effluent from the city of Katoomba, are mediated by inflow into the lake. Consequently, rainfall and runoff become the necessary intermediaries between events in the catchment (ie bushfires, agricultural practice) and the lake. In a region characterised by variable stream flow, this makes it difficult to find simple, direct relations between catchment events and the behaviour of the lake. This is clearly demonstrated by the failure to find any relationship between the application of fertilizer in the Shires surrounding and including the catchment of this lake, and the concentration of phosphorus at site 3D. Events, that might predispose the lake to abnormally large inputs of phosphorus, such as a bushfire, or the threatened removal of the superphosphate bounty, may have no noticeable effect whatever, unless there is rainfall sufficient to transport the available nutrients to the lake. Against a background of quite variable inflow, the delineation of trends, such as that towards nutrient enrichment and eutrophication, may be very difficult in Lake Burragorang.

Two approaches might be adopted to the assessment of any eutrophicating trends in Lake Burragorang. The first involves relying on data for dry years, to indicate any permanent change. This approach may be insensitive to the early stages of enrichment, because it uses data from the periods in which the link between the catchment, and behaviour in the lake, is weakest. The second would rely on the use of the relationships which have been found from the present data set, to give a picture of normal variation, against which to compare any future behaviour. Chlorophyll-a concentration at site 3D, may be approximately corrected for inflow by coupling the inflow-phosphorus and phosphorus-chlorophyll regressions given in Chapter 7, providing a rough estimate of the expected chlorophyll-a concentration for a given annual inflow. Comparison of expected and recorded concentrations of chlorophyll-a may be used to detect any trend towards enrichment of the lake.

Interestingly, under the likely operating conditions of the link between the Shoalhaven Scheme and Lake Burragorang (see Petrie and Smalls 1981), nutrient enriched inflow into Lake Burragorang will occur during dry years, and possibly in the spring months prior to the summer period of greatest water demand (and evaporative loss). In this case, the biological effect on Lake Burragorang may be readily recognised using the first of the above approaches; the more easily because the comparison would have the minimum possible interference from events in Lake Burragorang's catchment.

COMMENTS ON MANAGEMENT OF LAKE BURRAGORANG

Lake Burragorang is best considered mesotrophic, on the basis of chlorophyll and phosphorus concentrations (cf. Cullen and Rosich 1979), although it occupies the boundary between mesotrophy and oligotrophy in periods of low inflow. The lake, presents few of the classic water quality problems associated with the seasonal cycle of thermal stratification, and the attendant depletion of oxygen in the hypolimnion. Anoxia was recorded, at site 3D (outlet), only once in the 20 year study period, so that the marked deterioration of hypolimnetic water, with H_2S production, and regeneration of sedimented nutrients, is not a problem, as it has been for downstream Prospect Reservoir (Petrie and Smalls 1981). The unpredictable, and sometimes massive inflows, represent the greatest threat to the quality of water withdrawn from Lake Burragorang. This unpredictability means that management must respond quickly to sporadic events, and cannot reliably expect a certain water quality on a seasonal basis. However, because Lake Burragorang is thermally stratified for most of the year, there is usually vertical heterogeneity in the water column, and hence the potential for active management employing the selective withdrawal capabilities of the offtake structures at Warragamba Dam.

Although it is not possible to predict when an inflow will occur, it is possible to predict some of the effects of an inflow when one does occur. Turbidity is used here as a cardinal parameter in water quality considerations because (a) it is an obvious manifestation of an inflow, and may be used diagnostically in this regard, and (b) apart from its intrinsic nuisance value (e.g. aesthetic, chlorine demand) it is correlated with other nuisances, such as total iron and phosphorus.

A first step is to isolate periods with the greatest potential for water quality problems. The preceeding work facilitates this, with respect to the seasonal cycle of inflow type, and the related data on the turbidity of inflows. The following points may be useful:-

1. The prediction of seasonal interchange between interflow and underflow is possible from continuous monitoring of temperatures at four sites only - inflow of the Cox and Wollondilly Rivers plus surface and deep-water temperature at site 3D.
2. Turbidity, affecting the whole water column and therefore removing any selective withdrawal option, will be worst in the instance of floods that occur immediately before the period of maximum vertical mixing (July - August). This applies to either interflow or underflow, although higher turbidity is usually associated with the latter type, and is log-linearly related to the volume of underflows. June is a

particularly critical month in that (a) both interflows and underflows may occur, (b) either one at this time will elevate turbidity in the whole water column at the period of maximum mixing. This will increase nutrient levels in the near surface layers, and lead to algal blooms following re-establishment of the thermocline (e.g. 1974, 1975, 1978).

3. An inflow in August, when underflows are the only type of inflow that affects site 3D, is likely to be confined by advective stabilisation to the hypolimnion. Consequently, regardless of the relatively high turbidity, there remains an option to make use of selective withdrawal if the surface waters are of acceptable biological water quality. Further, this "locking up" of the newly influent water by advective stabilisation probably also reduces the transfer of particulate associated nutrients to the epilimnion at a time when such an injection of nutrients would result in a spring algal bloom. Insofar as the HEPS offtake is implicated in advective stabilisation, it therefore represents active management of the lake to reduce nutrient supply to the epilimnion.

4. Another potentially active management role of the HEPS, is the effect of its use on the thermal structure of the lake, particularly in the first four to five months of the year. Operation of the HEPS, at this time, will tend to draw the thermocline down towards the offtake level, effectively channeling interflows into the region where they might be directly withdrawn, at least in part. This positive effect must, however, be balanced against the tendency for the HEPS to draw less well oxygenated water from shallower sites, into the vicinity of the water supply offtake.

5. Since Prospect Reservoir draws a large proportion of its water from Lake Burragorang (Bowen and Smalls 1980), water quality at site 3D is

not the only concern. There are implications for the downstream storage. At the time of a major inflow the best quality water is near the surface and a logical first response is to take water from that zone rather than to shut abruptly the offtake to Prospect. Fig. 8.1 indicates that as recently as 1978 sub-optimal water was supplied to Prospect following the March interflow (as indicated by turbidity). This strategy avoids the introduction of nutrients and sediment to Prospect Reservoir. It is assumed that the resultant seeding-in of algae is unimportant because the reservoir is capable of seeding itself.

FUTURE WORK

The present analysis indicates some potentially fruitful pieces of research, directed at improving the existing data set, and extending the present analysis. These are listed below:-

1. The present data set might be considerably improved by experimentation designed to provide continuity for some parameters. Turbidity, measured by the old Hellige visual comparator, might be simply related to the new automated method, although the comparative insensitivity of the old method would have to be imposed on the newer data. An experiment comparing the two methods might use bottom mud collected at site 3D, although this will probably have changed somewhat in its light scattering properties after deposition. Alternatively, it may be necessary to await a turbid inflow, sample it, and concentrate the sample to acheive the full range of turbidity.

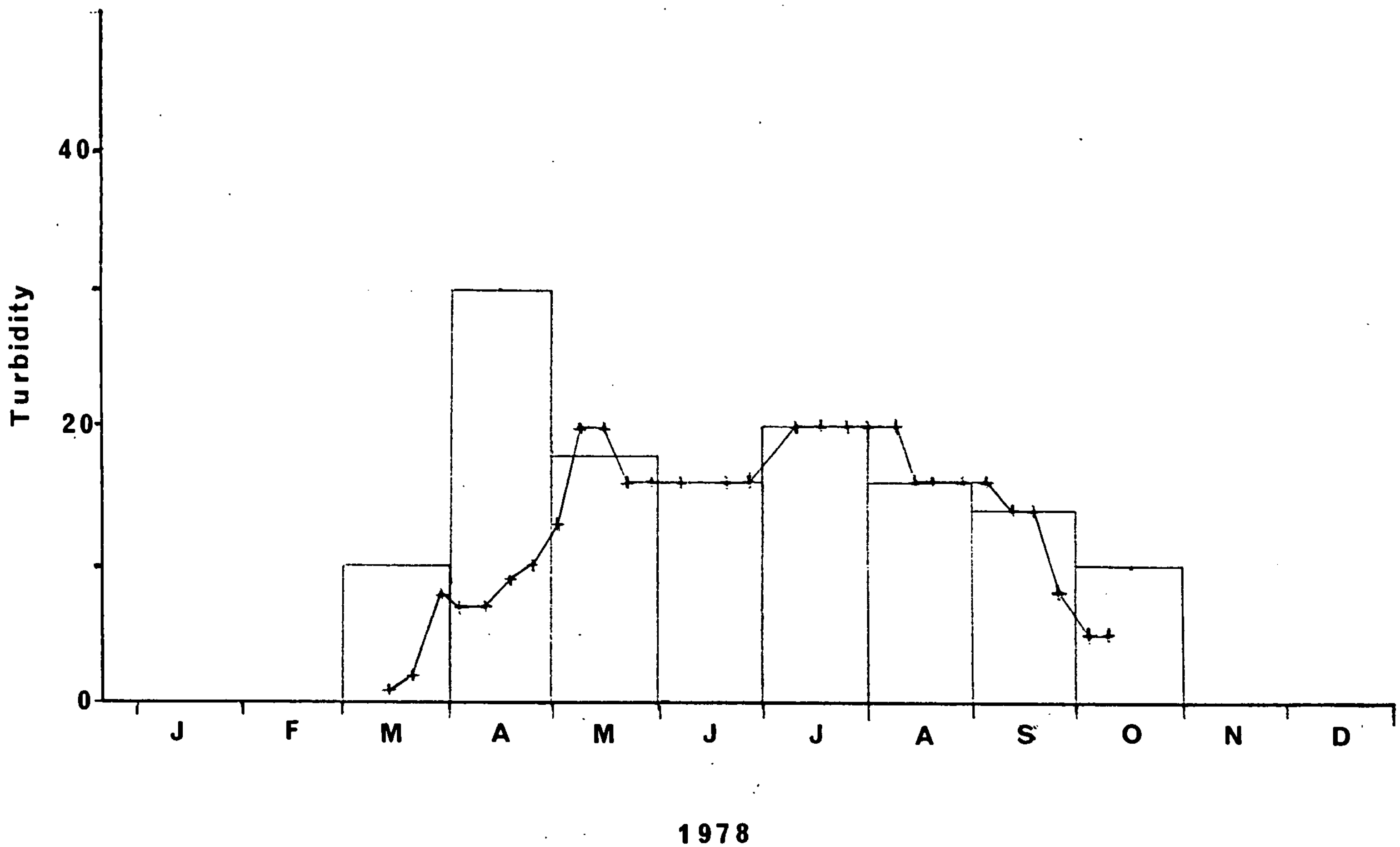
   The data for nitrate could also benefit from simple experimentation to ascertain whether or not the old and new methods can be inter-calibrated.
2. The successful comparison of WET and DRY years should be extended,

FIGURE 8.1

The monthly mean turbidity of water withdrawn through the variable offtake (c. 0 - 60 m) for supply to Prospect Reservoir, is shown as a bar graph for the period from March to October of 1978. The records of near surface turbidity, for the same period, are plotted as a line (+). The comparison indicates that water subtracted for supply to Prospect was of lower quality (based on turbidity) than was available, at least in March and April.

Symbols:-

Turbidity within 6 m of the surface (+)


Turbidity
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for instance in relation to the data for algal cell counts. Also, the more detailed thermal data collected by the Biology Section personnel could be analysed in this manner, and data for thermal eddy conductivity included.

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APPENDICES

APPENDIX 1

This appendix contains the instruction manual written to accompany a modified version of LIMNO, a program written by D. H. Merritt to calculate lake heat content, Schmidt stability, and Birgean wind work.

The equations used in the program, and some of the commentary about the calculated quantities, are relevant to the present work because the program was used in the analysis of Lake Burragorang's thermal stratification in Part 2 (Chapters 3 - 5) of the thesis. The program was also used extensively in the analysis of Deep Lake (Antarctica), which is the subject of a paper included in Appendix 3.

LIMNO/2 (November 1983)

J.M. Ferris

A modified version of LIMNO (by D.H. Merritt)

A BASIC program for calculation of whole lake stability, heat content and volume weighted averages of oxygen and salinity.

CONTENTS

LIMNO/2:

Introduction:

This is a modified version of LIMNO, an interactive program written (in BASIC) by D.H. Merritt (written: 1975). The original program is referred to in

Johnson, N. M., Eaton, J. S., and Richey, J. E. (1978). Analysis of five North American lake ecosystems. II. Thermal energy and mechanical stability. *Verh. Internat. Verein. Limnol.* **20**, 562-7.

and a listing or punched tape copy of the program may be obtained by writing to Dr N. M. Johnson, Earth Sciences Dept., Dartmouth College, Hanover, N.H. 03755, U.S.A.

In its original from LIMNO calculates for the whole lake:

1. Heat content (cal cm^{-2}) also for the whole lake volume (calories) and ice sheet (calories).
2. Volume weighted average temperature (°C).
3. Volume weighted average salinity (ppm).
4. Volume weighted average oxygen concentration (ppm).
5. Depth of the centre of density (metres) sensu Idso (1973).
6. Idso-modified Schmidt Stability (gm-cm cm^{-2}).
7. Birgean Wind-Work (gm-cm cm^{-2}).

The program gives the same basic physico-chemical averages for a 1 cm x 1 cm column in the lake.

The program also prints out depth profiles (at 1 metre intervals) of the following parameters:

1. Temperature (°C).
2. Salinity (ppm) or Conductivity (µmhos).
3. Density (gm cm^{-3}).
4. Oxygen concentration (ppm) and % saturation.
5. Idso-modified Schmidt Stability (gm-cm cm^{-2}).
6. Birgean Wind-Work (gm-cm cm^{-2}).

LIMNO/2 has somewhat extended capabilities:-

1. The Walker-modified Schmidt stability (Walker 1974), for both "Open" and "Closed" meromictic lakes (Sensu Hutchinson, Vol 1 1957) is calculated in profile and as a whole lake total.
2. LIMNO/2 will accept directly measured densities from which to calculate the lake stability. This facilitates its use with saline lakes.
3. The initial condition for the calculation of Birgean Wind-Work can now be specified in the data set, which generalises the calculation to include lakes outside the cool temperate zone (i.e. warm monomictic lakes).
4. Two external data files are used, instead of data statements within the program. This streamlines data handling.

Data Files:

Two data files, external to the program, are used. DYDATA, contains physico-chemical profiles for a single sampling, and is mandatory. HYPSOG contains lake hypsography, without it, LIMNO/2 will printout a limited amount of data pertaining to a 1 cm x 1 cm column of water.

The following, is a facsimile of printout, from the program if you answer "YES" to the question "Do you need help?". The instructions include the original data set for Mirror Lake (from LIMNO). Examples of data files for Mirror Lake, Deep Lake (Antartica), and Lake Burragorang (N.S.W.) are given in Figures 1 a-b.

```
TWO EXTERNAL DATA FILES ARE REQUIRED ; HYPSOG AND DYDATA
 IN THE FOLLOWING EXAMPLES LINES BEGINNING WITH A 'C'
ARE COMMENTS AND SHOULD NOT APPEAR IN AN ACTUAL DATA FILE.
DATA FILES SHOULD NOT HAVE 'BASIC' LINE NOS. EITHER, AS
THESE MAY BE REGARDED AS DATA !!
 NOTE : ALL STRINGS IN THE DATA FILE SHOULD BE IN
DOUBLE QUOTES, INSTEAD OF THE SINGLE ONES USED HERE !!

   DATA FILE 'HYPSOG'
C LAKE IDENTIFICATION STRING, FOR OUTPUT
 '*** HYPSOGRAPHY FOR MIRROR LAKE ***'
C MAX. LAKE DEPTH (>= ANY INDIVIDUAL DAILY DEPTH), AND
C ELEVATION OF THE LAKE BOTTOM ABOVE M.S.L. (metres)
11.1,200
C LIST OF AREAS WITHIN CONTOURS AT 1 METRE INTERVALS FROM
C TOP TO BOTTOM (GIVE EXACT BOTTOM AREA). (metres 2)
C NB: @ DENOTING '10 TO THE POWER OF' MAY BE DIFFERENT AT
C YOUR FACILITY.
15@4,13.6@4,12.4@4,11.5@4,10.5@4,9.86@4,8.96@4,6.79@4
3.21@4,1.61@4,0.609@4,0.1@4,0

   DATA FILE 'DYDATA'
C SAMPLE DAY I.D. STRING (FOR OUTPUT)
 'MIRROR LAKE 19 AUGUST 1969'
c LAKE DEPTH ON THE DAY (<= MAX. LAKE DEPTH IN HYPSOG),
C AND THE DESIRED PRINTOUT INTERVAL (METRES, >= 0.1)
C DEFAULT IS A PROFILE AT 1 METRE INTERVALS
11.0,0.5
C THICKNESS OF ICE (0.0 IF NO ICE : ACCURACY TO 0.01 M)
0
C  TEMP., CHEMICAL, D.O., AND DENSITY FOLLOW THE SAME
C  THREE-PART INPUT STRUCTURE.
C NUMBER OF DEPTHS SAMPLED FOR TEMPERATURE
11
C LIST OF SAMPLE DEPTHS (MAY OCCUPY > 1 LINE)
0,1,2,3,4,5,6,7,8,9,10
C LIST OF TEMPS. AT SPECIFIED DEPTHS
25,25,25,24.13,21.9,17.72,13.74,10.52,9.34,8.75,8.06
C NUMBER OF CHEMICAL SAMPLE DEPTHS (COND. OR SALINITY)
```

```
6
C TWO STRINGS, 'COND' OR 'SALINITY' AND 'UMHOS' or 'PPM'
'COND','UMHOS'
C LIST OF SAMPLE DEPTHS
0,2,4,6,8,10
C LIST OF CHEMICAL MEASUREMENTS
20.5,21.9,22.5,25.3,26
C DATA FOR OXYGEN PROFILE (PPM), as for TEMP.
11
0,1,2,3,4,5,6,7,8,9,10
7.35,7.44,7.46,7.54,8.3,7.62,6.61,1.52,0.18,0.1,0
C NB: IF A PARAMETER WAS NOT MEASURED (IE. OXYGEN)
C THEN THE NO. OF SAMPLE DEPTHS = 0, AND MOVE ON TO THE
C NEXT PROFILE.
C A PROFILE OF DIRECTLY MEASURED DENSITIES MAY BE PUT
C IN HERE, USING THE SAME THREE-PART STRUCTURE. IF THIS
C DATA DOES NOT EXIST THEN PUT NOTHING AT ALL; ONLY
C ANSWER THE APPROPRIATE INTERACTIVE QUESTION 'NO'.
C SECTION FOR SPECIFYING BIRGEAN INITIAL DENSITY VALUE.
C STRING(S), EITHER 'DENS' OR 'TEMP' FOLLOWED BY 'COND'
C OR 'SALINITY': (NB: IF 'TEMP' IS USED THEN A SECOND
C STRING MUST FOLLOW, '' WILL DO IF NEITHER OF THE OTHER
C OPTIONS IS TO BE USED.)
'TEMP','COND'
C ONE OR TWO NUMBERS CORRESPONDING TO THE STRING(S)
4.0,26
```

FIGURE 1a: HYPSOG files for Mirror Lake, Deep Lake, and Lake Burragorang, elevation and maximum depth (metres), area data (m^2).

```
100   "*** HYPSOGRAPHY FOR MIRROR LAKE ***"
200 11.1,200
300 15@4,13.6@4,12.4@4,11.5@4,10.5@4,9.86@4,8.96@4,6.79@4
400 3.21@4,1.61@4,.609@4,.1@4,0
#
```

```
100  *** HYPSOGRAPHY FOR DEEP LAKE (ANTARCTICA) ***
200 36.0,-86.0
300 635.227@3,613.382@3,591.537@3,569.693@3,547.848@3,526.003@3
400 515.990@3,505.977@3,495.963@3,485.950@3,475.937@3,466.538@3
500 457.139@3,447.741@3,438.342@3,428.943@3,415.640@3,402.337@3
600 389.034@3,375.731@3,362.428@3,344.314@3,326.200@3,308.086@3
700 289.972@3,271.858@3,252.946@3,234.035@3,215.123@3,196.212@3
800 177.300@3,149.188@3,121.075@3,92.963@3,64.850@3,36.738@3
900 0.446@3
#
```

```
100  *** HYPSOGRAPHY FOR LAKE BURRAGORANG ***
200 105.0,11.7
300 7500@4,7338@4,7172@4,7009@4,6847@4,6682@4,6518@4,6356@4
400 6197@4,6041@4,5889@4,5739@4,5591@4,5442@4,5291@4,5137@4
500 4987@4,4842@4,4702@4,4562@4,4421@4,4279@4,4141@4,4007@4
600 3878@4,3754@4,3634@4,3520@4,3404@4,3285@4,3159@4,3038@4
700 2924@4,2821@4,2717@4,2609@4,2496@4,2382@4,2270@4,2157@4
800 2048@4,1948@4,1856@4,1768@4,1749@4,1665@4,1583@4,1503@4
900 1426@4,1350@4,1277@4,1205@4,1136@4,1069@4,1004@4,942@4
1000 881@4,822@4,766@4,712@4,660@4,610@4,562@4,516@4,473@4
1100 431@4,392@4,355@4,320@4,287@4,256@4,227@4,201@4,176@4
1200 154@4,134@4,116@4,100@4,86@4,74@4,65@4,58@4,52@4,49@4
1300 48@4,46@4,44@4,41@4,39@4,37@4,34@4,32@4,30@4,27@4,25@4
1400 23@4,21@4,18@4,16@4,14@4,11@4,9@4,7@4,5@4,2@4,0
#
```

FIGURE 1b: DYDATA files. Temperature (°C), Density (gm cm^3).

```
100  "MIRROR LAKE 19 AUGUST 1969"
200 11.0,1.0
300 0
400 11
500 0,1,2,3,4,5,6,7,8,9,10
600 25,25,25,24.13,21.9,17.72,13.74,10.52,9.34,8.75,8.06
700 6
800 "COND","UMHOS"
900 0,2,4,6,8,10
1000 20.5,21,21.9,22.5,25.3,26
1100 11
1200 0,1,2,3,4,5,6,7,8,9,10
1300 7.35,7.44,7.46,7.54,8.3,7.62,6.61,1.52,.18,.1,0
```

```
100  "DEEP LAKE  26 JAN 1977"
200 35.8,2
300 0
400 21
500 0,4,7,8,9,10,11,12,13,14,15,16,17,18,19,20,21,25
600 30,33,34
700 7.5,7.5,7.4,7.3,5.2,-0.9,-3.7,-5.5,-6.5,-7.9,-8.8
800 -10.4,-11.9,-13.4,-14.3,-14.6,-15.0,-15.5,-15.7
900 -15.7,-15.6
1000 5
1100 "SALINITY","PPM"
1200 0,9,16,20,34
1300 224120,224120,223920,224900,224900
1400 0
1500 21
1600 0,4,5,8,9,10,11,12,13,14,15,16,17,18,19,20
1700 21,25,30,33,34
1800 1.1807,1.1807,1.1808,1.1808,1.1817,1.1844,1.1856,1.1864
1900 1.1868,1.1874,1.1877,1.1884,1.1894,1.1903,1.1910,1.1913
2000 1.1915,1.1917,1.1918,1.1918,1.1917
2100 "DENS"
2200 1.1921
```

```
100  "L. BURRAGORANG 3D 18 AUG 1980"
200 96.7,6
300 0
400 10
500 0,6,12,18,24,30,36,48,60,72
600 12.8,12.8,12.8,12.8,12.8,12.8,12.8,12.8,12.8,12.8
700 1
800 "COND","UMHOS"
900 0
1000 190
1100 10
1200 0,6,12,18,24,30,36,48,60,72
1300 8.7,8.7,8.4,8.5,8.6,8.4,8.5,8.4,8.4,8.0
1400 "TEMP","COND"
1500 12.8,190
*
```

Output:

Figures 2a-c show the LIMNO/2 output from the three chosen examples. These data files illustrate the two modes of density input, and the specification of a Birgean initial condition, other than isothermy at 4°C.

Mirror Lake, 19 AUGUST 1969:

Density is calculated from the temperature and conductivity profiles. The initial density for Birgean Wind-Work is assumed to be 1 gm cm^{-3} (density of fresh water at approximately 4°C)

Deep Lake, 26 JANUARY 1977:

Because the lake is hypersaline (c. 10 x sea-water) and sub-zero temperatures are common, the density/temperature relationship employed by LIMNO/2 is inappropriate. Density is input directly, and the initial density for Birgean Wind-work is specified as 1.1921 gm cm^{-3}.

A slight density anomaly is also found in this data set, and LIMNO/2 asks whether it is to be left in, or removed. In this instance removal involves declaring the water column below 33 metres to have a density of 1.1918 gm cm^{-3} rather than 1.1917 gm cm^{-3}. These anomalies are relatively common near the lake surface during the cooling phase.

Lake Burragorang, 18 AUGUST 1980:

A winter profile for this warm monomictic lake, in which temperature and conductivity are used to calculate the density profile, and the initial density for Birgean Wind-Work. Unless density is corrected for volumetric contraction due to depth, the lake is declared non-stratified. In LIMNO/2 this

FIGURE 2a: LIMNO/2 output for Mirror lake (19 August 1969).

```
DO YOU NEED HELP ??
NO
IS YOUR LAKE'S AREA DATA IN FILE HYPSOG ?
YES
DO YOU WANT A PROFILE PRINTED ?
YES
DO YOU WANT DATA PRINTED AS DEPTH OR ELEV ?
DEPTH
ARE YOU USING PRE-CALCULATED DENSITIES ?
NO
CORRECT DENSITY FOR VOLUMETRIC CONTRACTION DUE TO DEPTH ?
NO
DO YOU WANT DENSITY CORRECTED FOR SALINITY ?
YES
DO YOU WANT O2 % SATURATION DEPTH CORRECTED ?
NO
IS THE LAKE *OPEN* OR *CLOSED* (WALKER-SCHMIDT CALC) ?
CLOSED


*** HYPSOGRAPHY FOR MIRROR LAKE ***

DATA FOR MIRROR LAKE 19 AUGUST 1969
```

DEPTH METRES	TEMP DEG(C)	COND UMHOS	DENSITY GM/CM**3	OYGEN PPM	%SAT	SCHMIDT IDSO :	WALKER	BIRGEAN (GM-CM/CM**2)
.0	25.00	20.5	.997093	7.35	90.17	3.68	.00	.00
1.0	25.00	20.8	.997093	7.44	91.28	2.53	.00	2.62
2.0	25.00	21.0	.997094	7.46	91.52	1.58	.00	4.80
3.0	24.13	21.5	.997313	7.54	91.09	.59	.22	6.16
4.0	21.90	21.9	.997840	8.30	96.30	.02	.91	6.05
5.0	17.72	22.2	.998692	7.62	81.58	.39	2.25	4.28
6.0	13.74	22.5	.999323	6.61	65.12	1.44	3.22	2.36
7.0	10.52	23.9	.999695	1.52	13.92	2.05	3.10	.90
8.0	9.34	25.3	.999798	.18	1.60	1.39	1.76	.32
9.0	8.75	25.7	.999842	.10	.88	.88	1.03	.14
10.0	8.06	26.0	.999887	.00	.00	.40	.44	.04
11.0	8.06	26.0	.999887	.00	.00	.00	.00	.00

```
DATA FOR THE TOTAL LAKE
THE TOTAL HEAT CONTENT OF THE LAKE IS  1.7446283+13 CALORIES
HEAT CONTENT OF THE LAKE IS  11740.43 CAL/CM**2
AVERAGE TEMPERATURE IS  20.501
AVERAGE SALT CONC IS  14.954  PPM
TOTAL AMOUNT OF SALT IS  12725896.433  GRAMS
AVERAGE O2 CONTENT IS  6.605  PPM
TOTAL MOLES O2 IS  175654.87192
THE CENTRE OF DENSITY IS  4.154  METRES BELOW THE SURFACE
MEAN DENSITY OF THE LAKE IS  0.9979831  GM/CM**3
THE IDSO - SCHMIDT STABILITY IS    131.73  GM-CM/CM**2
EPILIMNION IS ** CLOSED **
THE WALKER - SCHMIDT STABILITY IS  130.36  GM-CM/CM**2

         THE BIRGEAN WIND WORK IS  278.16  GM-CM/CM*2
INITIAL DENSITY FOR BIRGEAN WORK =  1  GM/CM**3

LAKE SURFACE AREA =  148600  SQ. METRES
TOTAL LAKE VOLUME =  851010  M.**3

DATA FOR A CM X CM COLUMN
HEAT CONTENT OF WATER COLUMN IS  18234.3  CALORIES
AVERAGE SALT CONC IS 15.932536365  PPM
AVERAGE O2 CONTENT IS  4.6193181818  PPM
TOTAL MOLES O2 IN CM**2 COLUMN IS  1.58789063-4  MOLES
```

FIGURE 2b: LIMNO/2 output for Deep Lake (26 January 1977).

```
DO YOU WANT A PROFILE PRINTED ?
?YES
DO YOU WANT DATA PRINTED AS DEPTH OR ELEV ?
?DEPTH
ARE YOU USING PRE-CALCULATED DENSITIES ?
?YES
CORRECT DENSITY FOR VOLUMETRIC CONTRACTION DUE TO DEPTH ?
?NO
DO YOU WANT O2 % SATURATION DEPTH CORRECTED ?
?NO
IS THE LAKE *OPEN* OR *CLOSED* (WALKER-SCHMIDT CALC) ?
?CLOSED

ANOMALOUS DENSITY FOUND AT  33.1  METRES
DO YOU WANT TO REMOVE THE ANOMALY ?
?NO

*** HYPSOGRAPHY FOR DEEP LAKE (ANTARCTICA) ***

DATA FOR DEEP LAKE  26 JAN 1977
```

DEPTH METRES	TEMP DEG(C)	CHEM PPM	DENSITY GM/CM**3	OXYGEN PPM	%SAT	SCHMIDT IDSO	: WALKER	BIRGEAN (GM-CM/CM**2)
.0	7.50	224120.0	1.180700	.00	.00	59.48	.00	.00
2.0	7.50	224120.0	1.180700	.00	.00	45.65	.00	21.18
4.0	7.50	224120.0	1.180700	.00	.00	33.26	.00	39.21
6.0	7.43	224120.0	1.180800	.00	.00	22.54	.19	55.19
8.0	7.30	224120.0	1.180800	.00	.00	13.65	.19	70.71
10.0	00-.0	224091.4	1.184400	.00	.00	1.61	13.37	57.80
12.0	-5.50	224034.3	1.186400	.00	.00	.20	21.58	49.31
14.0	-7.90	223977.1	1.187400	.00	.00	2.65	26.12	45.48
16.0	-10.40	223920.0	1.188400	.00	.00	7.43	30.75	38.69
18.0	-13.40	224410.0	1.190300	.00	.00	17.64	40.40	19.81
20.0	-14.60	224900.0	1.191300	.00	.00	26.21	44.64	9.08
22.0	-15.13	224900.0	1.191550	.00	.00	30.39	43.17	6.17
24.0	-15.38	224900.0	1.191650	.00	.00	32.63	40.59	4.89
26.0	-15.54	224900.0	1.191720	.00	.00	33.28	37.41	3.89
28.0	-15.62	224900.0	1.191760	.00	.00	32.30	33.60	3.17
30.0	-15.70	224900.0	1.191800	.00	.00	29.49	28.95	2.43
32.0	-15.70	224900.0	1.191800	.00	.00	21.88	20.67	1.74
34.0	-15.60	224900.0	1.191700	.00	.00	11.96	11.09	1.25
35.8	-15.60	224900.0	1.191700	.00	.00	.00	.00	.00

```
DATA FOR THE TOTAL LAKE
THE TOTAL HEAT CONTENT OF THE LAKE IS -4.8314599+13 CALORIES
HEAT CONTENT OF THE LAKE IS -7658.55 CAL/CM**2
AVERAGE TEMPERATURE IS -3.779
AVERAGE SALINITY CONC IS  224323.177  PPM
TOTAL AMOUNT OF SALINITY IS  2.8678016+12  GRAMS
THE CENTRE OF DENSITY IS  11.405  METRES BELOW THE SURFACE
MEAN DENSITY OF THE LAKE IS  1.1859243000  G/CM**3
*** A MAXIMUM DENSITY GREATER THAN 1 HAS BEEN USED ***

THE IDSO - SCHMIDT STABILITY IS    7865.76  GM-CM/CM**2
EPILIMNION IS ** CLOSED **
THE WALKER - SCHMIDT STABILITY IS  7838.12  GM-CM/CM**2

         THE BIRGEAN WIND WORK IS  8687.65  GM-CM/CM**2
INITIAL DENSITY FOR BIRGEAN WORK =  1.1921  GM/CM**3

LAKE SURFACE AREA =  630858  SQ. METRES
TOTAL LAKE VOLUME =  12784241.000  M.**3

DATA FOR A CM X CM COLUMN
HEAT CONTENT OF WATER COLUMN IS -26708.5  CALORIES
AVERAGE SALINITY CONC IS 225104.21780  PPM
```

FIGURE 2c: LIMNO/2 output for Lake Burragorang (18 August 1980).

```
DO YOU NEED HELP ??
NO
IS YOUR LAKE'S AREA DATA IN FILE HYPSOG ?
YES
DO YOU WANT A PROFILE PRINTED ?
YES
DO YOU WANT DATA PRINTED AS DEPTH OR ELEV ?
DEPTH
ARE YOU USING PRE-CALCULATED DENSITIES ?
NO
CORRECT DENSITY FOR VOLUMETRIC CONTRACTION DUE TO DEPTH ?
NO
DO YOU WANT DENSITY CORRECTED FOR SALINITY ?
YES
DO YOU WANT O2 % SATURATION DEPTH CORRECTED ?
NO
IS THE LAKE *OPEN* OR *CLOSED* (WALKER-SCHMIDT CALC) ?
CLOSED

*** HYPSOGRAPHY FOR LAKE BURRAGORANG ***

DATA FOR L. BURRAGORANG 3D 18 AUG 1980
```

DEPTH METRES	TEMP DEG(C)	COND UMHOS	DENSITY GM/CM**3	OXYGEN PPM	%SAT	SCHMIDT IDSO	: WALKER	BIRGEAN (GM-CM/CM**2)
.0	12.80	190.0	.999536	8.70	82.94	.00	.00	.00
6.0	12.80	190.0	.999536	8.70	82.94	.00	.00	.00
12.0	12.80	190.0	.999536	8.40	80.08	.00	.00	.00
18.0	12.80	190.0	.999536	8.50	81.03	.00	.00	.00
24.0	12.80	190.0	.999536	8.60	81.99	.00	.00	.00
30.0	12.80	190.0	.999536	8.40	80.08	.00	.00	.00
36.0	12.80	190.0	.999536	8.50	81.03	.00	.00	.00
42.0	12.80	190.0	.999536	8.45	80.56	.00	.00	.00
48.0	12.80	190.0	.999536	8.40	80.08	.00	.00	.00
54.0	12.80	190.0	.999536	8.40	80.08	.00	.00	.00
60.0	12.80	190.0	.999536	8.40	80.05	.00	.00	.00
66.0	12.80	190.0	.999536	8.20	78.14	.00	.00	.00
72.0	12.80	190.0	.999536	8.00	76.27	.00	.00	.00
78.0	12.80	190.0	.999536	8.00	76.27	.00	.00	.00
84.0	12.80	190.0	.999536	8.00	76.27	.00	.00	.00
90.0	12.80	190.0	.999536	8.00	76.27	.00	.00	.00
96.0	12.80	190.0	.999536	8.00	76.27	.00	.00	.00
96.7	12.80	190.0	.999536	8.00	76.27	.00	.00	.00

```
DATA FOR THE TOTAL LAKE
THE TOTAL HEAT CONTENT OF THE LAKE IS  2.0218678+16 CALORIES
HEAT CONTENT OF THE LAKE IS  32874.83 CAL/CM**2
AVERAGE TEMPERATURE IS  12.8
AVERAGE SALT CONC IS  129.96  PPM
TOTAL AMOUNT OF SALT IS  205282762632  GRAMS
AVERAGE O2 CONTENT IS  8.534  PPM
TOTAL MOLES O2 IS  421278996.51

** THE LAKE IS NOT DENSITY STRATIFIED **

DATA FOR A CM X CM COLUMN
HEAT CONTENT OF WATER COLUMN IS  123904  CALORIES
AVERAGE SALT CONC IS 130.09439508  PPM
AVERAGE O2 CONTENT IS  8.337072562  PPM
TOTAL MOLES O2 IN CM**2 COLUMN IS  .00251935911  MOLES
```

declaration is printed when the density at the lake bottom (equivalent to density at deepest sampled depth) is less than, or equal to the volume weighted mean density of the lake.

Figure 2d shows the output from the original LIMNO, as implemented on the University of Tasmania Burroughs B6800 mainframe computer. The data is for Mirror Lake. LIMNO/2 gives slightly different results for the stability measures. The differences result mainly from a change in the value of the Lake surface area. Due to a peculiarity in incrementation of the counter in BASIC for/next loops, LIMNO uses a surface area 0.1 m above the actual lake surface (at least as implemented on our system). This led to the situation where LIMNO was unable to accept a daily depth (in DYDATA) equal to the maximum lake depth (in HYPSOG).

FIGURE 2d: Output from the original LIMNO, for Mirror Lake (19 August 1969)

```
LIST THIS PROGRAM FOR INSTRUCTIONS

IS YOUR RESERVOIRS DATA IN THIS PROGRAM ?
#?
?YES
DO YOU WANT A PROFILE PRINTED ?
?YES
DO YOU WANT DATA PRINTED AS DEPTH OR ELEV ?
?DEPTH
DO YOU WANT DENSITY CORRECTED FOR VOLUMETRIC CONTRACTION ?
DUE TO DEPTH
?NO
DO YOU WANT DENSITY CORRECTED FOR SALINITY ?
?YES
DO YOU WANT O2 SATURATION DEPTH CORRECTED
?NO
*** HYPSOGRAPHY IS FOR MIRROR LAKE ***

DATA FOR MIRROR LAKE FOR19 AUGUST 1969


DEPTH   TEMP  COND  DENSITY     OXYGEN    SCHMIDT  BIRGEAN
METRES  DEG(C) UMHOS GM/CM**3   PPM   %SAT      GM-CM/CM**2

-    .0 -25.0 -20.5 - .997093 - 7.35 -90.17  - 3.64  -   .00
-   1.0 -25.0 -20.8 - .997093 - 7.44 -91.28  - 2.51  - 2.60
-   2.0 -25.0 -21.0 - .997094 - 7.46 -91.52  - 1.57  - 4.75
-   3.0 -24.1 -21.5 - .997313 - 7.54 -91.09  -  .59  - 6.10
-   4.0 -21.9 -21.9 - .997840 - 8.30 -96.30  -  .02  - 5.99
-   5.0 -17.7 -22.2 - .998692 - 7.62 -81.58  -  .39  - 4.24
-   6.0 -13.7 -22.5 - .999323 - 6.61 -65.12  - 1.42  - 2.34
-   7.0 -10.5 -23.9 - .999694 - 1.52 -13.92  - 2.03  -   .89
-   8.0 - 9.3 -25.3 - .999798 -  .18 - 1.60  - 1.38  -   .32
-   9.0 - 8.8 -25.7 - .999842 -  .10 -   .88 -  .88  -   .14
-  10.0 - 8.1 -26.0 - .999887 -  .00 -   .00 -  .40  -   .04
-  11.0 - 8.1 -26.0 - .999887 -  .00 -   .00 -  .00  -   .00
-  11.0 - 8.1 -26.0 - .999887 -  .00 -   .00 -  .00  -   .00

DATA FOR THE TOTAL LAKE
THE HEAT CONTENT OF THE TOTAL LAKE IS 1.7446283+13 CALORIES
HEAT BUDGET OF THE LAKE IS 11630.855384 CAL/CM**2
AVERAGE TEMPERATURE IS 20.500679282
AVERAGE SALT CONC IS  14.953874142  PPM
TOTAL AMOUNT OF SALT IS 12725896.433  GRAMS
AVERAGE O2 CONTENT IS  6.6050409531  PPM
TOTAL MOLES O2 IS  175654.87192
THE CENTRE OF DENSITY IS  4.1541913066  METRES BELOW THE SURFACE
THE SCHMIDT STABILITY IS  130.48689826  GM-CM/CM**2
THE BIRGEAN STABILITY IS  275.59336654  GM-CM/CM**2

DATA FOR A CM X CM COUMN
HEAT CONTENT OF WATER COLUMN IS  18234.3  CALORIES
AVERAGE SALT CONC IS 15.932536365  PPM
AVERAGE O2 CONTENT IS  4.6193181818  PPM
TOTAL MOLES O2 IN CM**2 COLUMN IS  1.58789063-4  MOLES
```

Assumptions:

LIMNO/2 assumes :-

1. That a single temperature profile represents the whole lake. If this is unacceptable, a single profile averaged over several sample sites may be used.
2. That the temperature, or concentration for the deepest sampled depth can be extrapolated to the lake bottom.
3. That the centre of density (sensu Idso 1973) is unique in the water column. This is not always true, especially during the late cooling phase of a warm monomictic lake. If LIMNO/2 is asked not to correct such a density anomaly then the centre of density will be taken as the deepest point at which the volume weighted mean density occurs in the density profile.

Restrictions:

LIMNO/2 has the following restrictions:-

1. Data is linearly interpolated to provide an integration interval of 0.1 m. Several large data storage arrays are used (more than in LIMNO), and consequently running the program requires considerable memory space, and is presumably more expensive than if these arrays could be used more efficiently. The size of these arrays (declared at lines 300-320) is currently sufficient to handle a lake up to 120m deep. If much shallower lakes are under study the arrays dimensioned to (1200) may be reduced to 10 times the maximum lake depth, and those dimensioned to (120) may be reduced to the maximum anticipated number of sample depths.
2. For each parameter (except density, which is the subject of one of the interactive questions), at least a surface value must be given. All

depths should be specified to the nearest 0.1 m, except ice thickness which may be given to the nearest 0.01 m.

Program Transfer to non-Burroughs Computers:

Burroughs BASIC, is comparatively elementary, and should transfer readily to other installations. There are however certain language elements that may need to be changed. The following list is mainly a chronicle of things that had to be changed in the original program before it ran successfully on the B6800.

1. REM, is used here to denote a remark, an apostrophe may be required at your installation.
2. Relational expressions, (<=, <>) may be different (i.e. LE, NE).
3. Burroughs BASIC uses the ampersand "@" to represent "times 10 to the power" (i.e. 1@4 = 1 x 10^4) whereas LIMNO had "E".
4. Burroughs BASIC denotes exponentiation by "**" as opposed to " ↑ " in the original program.
5. The various characters appearing in Burroughs BASIC "PRINT USING" commands may differ at your installation.

The following is a list of BASIC line Nos. where the program elements, listed above, occur in LIMNO/2, outside of comments or strings.

REM	Lines 10-570, 630, 850, 870, 940, 990, 1010, 1030, 1070, 1130, 1170, 1440, 1580, 1630, 1770, 1830, 2150, 2390, 2550, 2570, 2590, 2610, 2660, 2670, 2680, 2710, 2780, 2820, 3050, 3070, 3090, 3150, 3170, 3200, 3240, 3250, 3270, 3290, 3310, 3400, 3440, 3460, 3490, 3750, 3770, 3820, 3900, 3960, 3990, 4090, 4120-4160, 4180, 4190, 4310, 4450, 4790, 5620
<=	700, 3720, 3790, 3980, 5280, 5340
>=	890
<>	1060, 1610, 1650, 2170, 2700, 3360, 3380, 3540, 3570, 3600, 3630,

3680, 4020, 5040

@ 1290, 1410, 1560, 1960, 2100, 2380, 2510, 3160, 3260, 4360, 4370, 4390, 4410, 5130, 5170, 5580, 5600

** 2620, 3180, 3410, 3470, 3580, 3760, 3970, 4100, 4260, 4290, 4460, 4470, 4480, 5130, 5170, 5220

PRINT USING images i.e. * and 'E occur only at lines 4690-4740

Magnetic Tape Library:

Three 1200 foot magnetic tapes have been recorded with triplicate copies of,

1. LIMNO/2: The most recent modification of the original LIMNO.
2. LIMNO/ANS: A version of LIMNO/2 which reads answers to the interactive questions from a third external data file "LIMNIN" (a data file of strings). This speeds things up somewhat.

and duplicate copies of,

1. LIMNO: The original program as implemented on the B6800 (contains data for Mirror Lake).
2. Three test files: one each from Mirror Lake (U.S.A.) Deep Lake (Antartica), and Lake Burragorang (N.S.W.). These files are for LIMNO/2 and LIMNO/ANS, and copies of the expected output appear in this manual.
3. LIMNIN: A file containing text strings answering the interactive questions that are asked by LIMNO/ANS. Note: this file is appropriate for the Mirror Lake and Lake Burragorang test data files only and not in the case of the Deep Lake data which uses directly determined density data.

Tapes have been written using TAS/TAPE/EXPORTER a utility program which uses a generalised format for transfer of files to non-Burroughs computers. Tapes are unlabelled, multi-file reels, with fixed length records, arranged in blocks of ten records, and using the ASCII character set. The following extract is from Uni. of Tas information bulletin (TAS/TAPE/EXPORTER: R.K. Allen 1982). Points of greatest relevance are marked with * or **.

DETAILS

1. The files to be copied to the tape are either explicitly named or implied by naming a directory, however ...
2. No codefiles and other Burroughs specific files are copied, in fact the only files that will be copied are symbolic files (which CANDE knows as SEQ, COBOL, FORTRAN, PASCAL, JOB, etc) and DATA files (DATA and CDATA). Unfortunately the program cannot distinguish between data files containing only character data and those containing machine-specific binary data. Files in the second category give gobbledy-gook when listed by CANDE, and included saved SPSS system files. They will be copied by the program but will be completely useless, perhaps unreadable, at the "foreign" installation. (SPSS users see note 11).

3.* Normal character files (including all symbolic files) will be written using the ASCII character set with the eighth bit zero.

4.* All symbolic files will be written as 80 character records, and so will 14-word DATA files (the program assumes these are created by the CANDE command MAKE filename DATA).

5.* The record length for all other files (i.e. CDATA and DATA files that are not 14-word) will be the same as that on the Burroughs disk. Hence printfiles, that is, program output directed to disk, may have record lengths of 120, 132, 133 or 138 characters depending on the program that created them. (The latter two lengths allow space for 132 columns plus FORTRAN carriage control.)

6.* Records are grouped into blocks. There are ten records per block unless this would give more than 2048 characters in a block, in which case a smaller blocking factor is used.

7.* Files on the tape are unlabelled and separated by single tapemarks; a double tapemark follows the last.

8.** The first file on the tape consists of 80 character records in 800 character blocks and contains a list of files written. For each file the following information is given: its number along the tape, its length in records, its record length (in characters), its blocksize (in characters), its type (as known to CANDE, e.g. SEQ, DATA), its Burroughs title (up to a maximum of 40 characters). The first file itself is shown with the title A/DIRECTORY.

9. The program produces a printer listing of the first file.

10.* By default the tape is written phase-encoded (PE) at a density of 1600 BPI. For NRZI encoding with 800 BPI give the following file equation immediately after the RUN command:

FILE OUTPUT(TAPE9);

If your computing installation uses a large (B5900/6700/7700 series) Burroughs computer then an alternate tape format is preferable and a fourth tape, with the alternate format, can be obtained from:-

Mr. H. R. Burton,
Antarctic Division,
Dept. of Science and Technology,
Channel Highway, Kingston,
Tasmania 7150 AUSTRALIA

Finally, it is particularly important that you return the tape as soon as possible.

Equations and Major Calculation Steps in LIMNO/2 (in order of appearance).

Note: The integration interval (dz) is a constant equal to 0.1 metres.

1. Volume, calculated from interpolated area data.

$V_z = (A_z + A_z + 0.1/2)\,dz$ (V_z is equivalent to $A_z dz$)

$$V_{tot} = \int_{z_0}^{z_m} V_z \qquad \text{(units : metres}^3\text{)}$$

2. Totals of Heat content, O_2 concentration, Salinity and Density are calculated (after interpolation to 0.1 m intervals) according to the following general formula.

$$\text{Total} = \int_{z_0}^{z_m} \text{Parameter} \times V_z \times \text{constant}$$

where the "constant" is to arrive at the desired units.

These totals may subsequently be expressed per unit surface area (i.e. Heat content), or per unit volume (i.e. average O_2 and Salinity).

3. Heat content of ice sheet.

$H_i = V_{tot,i} \times 0.917$ (density) $\times$ 79.6 (latent heat of fusion) $\times$ constant (to give correct units) (units : calories)

4. Conversion of conductivity to salinity (Conductivity assumed to be measured at 18°C).

Salinity (ppm) = Conductivity (µmhos) x 0.684

5. Conversion of depth to pressure.

$$P_z = z/10.17 \quad \text{(Units : atmospheres)}$$

6. Conversion of altitude (of lake surface) to pressure.

   P_0 = (758.39 - (0.08758 x Surface elevation AMSL))/760 (units: atms)

7. Density due to temperature (distilled water; Tilton and Taylor 1937).

   $p_{t,z} = 1 - (((t_z-3.9863)^2/508929.2) \times ((t_z+288-19414)/(t_z+68.12963)))$

   (units : gm cm^{-3})

8. Multiplier to correct for density increase with depth

   $M_{D,z} = 1/(1 - (P_z\ 0.00005))$ (0.00005 = coefficient of volumetric contraction)

9. Additive contribution, to density, due to salinity

   $p_{s,z}$ = Salinity x (8×10^{-7}) (units gm cm^{-3})

10. Final density term, at depth z, including contributions from temperature, salinity and depth.

    $p_z = (p_{t,z} M_{D,z}) + p_{s,z}$

11. % saturation of oxygen at depth z.

    Saturation (O_{sat}) at a given temperature and pressure (P), where P may include contributions from both depth and altitude, or just the latter.

    $O_{sat,z} = P_z/(0.06719 + (0.00209\ t_z))$ (units : ppm)

    $O_{\%sat,z} = (O_z/O_{sat,z}) \times 100$

12. Corrections of density anomalies (if requested)

    A loop compares adjacent density values (p_z) from the surface downwards.

    if $p_z > p_{z+0.1}$ then replace $p_{z+0.1}$ with p_z

    LIMNO/2 re-calculates the volume weighted mean density of the lake of such a correction is performed.

13. Centre of density ($z_{\bar{p}}$) (Idso-modification of Schmidt Stability).

The position ($z_{\bar{p}}$) of the volume weighted mean density in the stratified water column is found by a loop which searches from the lake bottom upwards. (units : metres)

14. Walker - Schmidt Stability : Calculation of mean density and centre of density of the downward mixing layer.

Depth of the volumetric centre of gravity ($z_{g,z}$) of water above depth z.

$$z_{g,z} = 1/V_{tot,z} \int_0^z z\, V_z \quad \text{(where } V_z \text{ is considered equivalent to } A_z dz\text{)}$$

Mixed density of water above z (metres)

$$p_{m,z} = 1/V_{tot,z} \int_0^z p_z\, V_z \quad \text{(where } V_z \text{ equivalent to } A_z dz\text{)}$$

15. Stability Calculations.

$$S_{I,z} = A_0^{-1} \int_{z_0}^{z_m} (z_{\bar{p}} - z)(\bar{p} - p_z)\, V_z \times \text{constant}$$

$$B_z = A_0^{-1} \int_{z_0}^{z_m} z\,(p_i - p_z)\, V_z \times \text{constant}$$

Walker - Schmidt Stability for a "Closed" lake.

$$S_{W,z}^{\text{closed}} = A_0^{-1} \int_{z_0}^{z_m} (z - z_{g,z})(p_z - p_{m,z})\, V_z \times \text{constant}$$

Walker-Schmidt Stability for an "Open" lake

$$S^{open}_{W,z} = A_0^{-1} \int_{z_0}^{z_m} (z - z_{g,z})(p_z - p_0)\, Vz \times \text{constant}$$

Totals of the above stability measures are accumulated:

$$\text{Stability Total} = \int_{z_0}^{z_m} \text{Stability}_z \quad (\text{i.e. } S_l = \int_{z_0}^{z_m} S_{l,z})$$

Symbols:

1. Area : A : metres^2

   A_0 = surface area of the lake.

   A_z = area at depth z.

   $A_{z0.1}$ = area at 0.1m below z.

2. Density : p : gm cm^{-3}

   $\bar{p}$ = volume weighted mean density of the lake.

   p_i = initial density for Birgean Wind Work.

   $p_{m,z}$ = mixture density of water above z. (Walker Stability).

   p_0 = density of surface water.

   $p_{s,z}$ = density, at depth z (m), due to salinity (additive).

   $p_{t,z}$ = density, at z, due to temperature (for distilled water).

   p_z = total density term for water at depth z.

   $p_{z+0.1}$ = total density term for water at z + 0.1 metres.

   Density correction for depth : M : none

   $M_{D,z}$ = multiplier to correct for density increase due to overlying z metre water column.

3. Depth : z : metres

d_z = integration interval, 0.1m in LIMNO/2.

z = depth (unspecified).

$z+0.1$ = $z + d_z$, 0.1m deeper than z.

$z_{g,z}$ = depth of volumetric centre of gravity above z (Walker Stability).

z_0 = lake surface.

z_m = maximum daily depth.

$z_{\bar{p}}$ = depth of $\bar{p}$ in stratified profile (Idso Stability).

4. Heat Content: H : calories

H_i = heat content of the ice sheet.

5. Oxygen : O : ppm

$O_{sat,z}$ = concentration equivalent to saturation of oxygen at z.

$O_{\%sat,z}$ = O_z as percentage of $O_{sat,z}$ (no units).

O_z = oxygen concentration at depth z.

6. Pressure : P : atmospheres

P_0 = pressure, due to altitude, at lake surface.

P_z = pressure due to weight of z metre water column.

7. Stability : S ; B : gm-cm cm^{-2}

$S_{I,z}$ = Idso-Schmidt stability at z.

$S_{W,z}^{closed}$ = Walker-Schmidt stability at z, for an ideal "closed" lake.

$S_{W,z}^{open}$ = as above, for an ideal "open" lake.

B_z = Birgean Wind-Work at depth z.

S_l : S_W^{closed} : S_W^{open} : B = whole lake totals.

8. Temperature : t : °C

t_z = temperature at z metres.

9. Volume : V : metres^3

V_{tot} = total lake volume.

$V_{tot,i}$ = total volume of 1 metre thick ice sheet.

$V_{tot,z}$ = total volume above z.

V_z = volume vector at depth z.

General comments:

1. The implementations of the Walker-modified Schmidt Stability is specifically suited to meromictic lakes or lakes in which mechanical stability is primarily due to factors other than temperature. If the epilimnion (mixolimnion) is declared "CLOSED" then the calculation yields a total lake stability very similar to the Idso-modified Schmidt Stability (in all instances we've tried). If the epilimnion is considered "OPEN" then the Walker-Schmidt Stability is increased. It should be noted that this assumption is only valid where the contribution of temperature to the overall stability is minimal. It would be an unusual lake where the conceptual mixing event led to holomixis at the temperature of the surface water, rather than at the volume weighted mean temperature of the lake.This would clearly violate Schmidt's definition of stability. "The amount of work needed to mix the entire body of water to uniform temperature without addition or subtraction

of heat" (Hutchinson, Vol 1 1957).

The difference between the Idso-Schmidt Stability and the Walker-Schmidt Stability for an "open" system should give a stability range corresponding to that between complete closure and a totally open system.

2. A more powerful version of LIMNO, by the original author, is discussed in Johnson and Merritt (1979). The newer version is apparently capable of accepting input from more than one site in the reservoir, and allows a Birgean initial profile, as opposed to a single value, to be specified for a monomictic lake.

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APPENDIX 2

This appendix contains examples of temperature (°C) and dissolved oxygen (mg l^{-1}) profiles, for WET (1963, 1974, 1976, 1978), and DRY (1965, 1968, 1979, 1980) year groups. A representative profile, for each of the two parameters, taken from the sample day closest to the 15th of each month, is plotted. One years data (12 profile pairs) is given on each of the following 8 pages.

On each page, the plots are ordered as follows:-

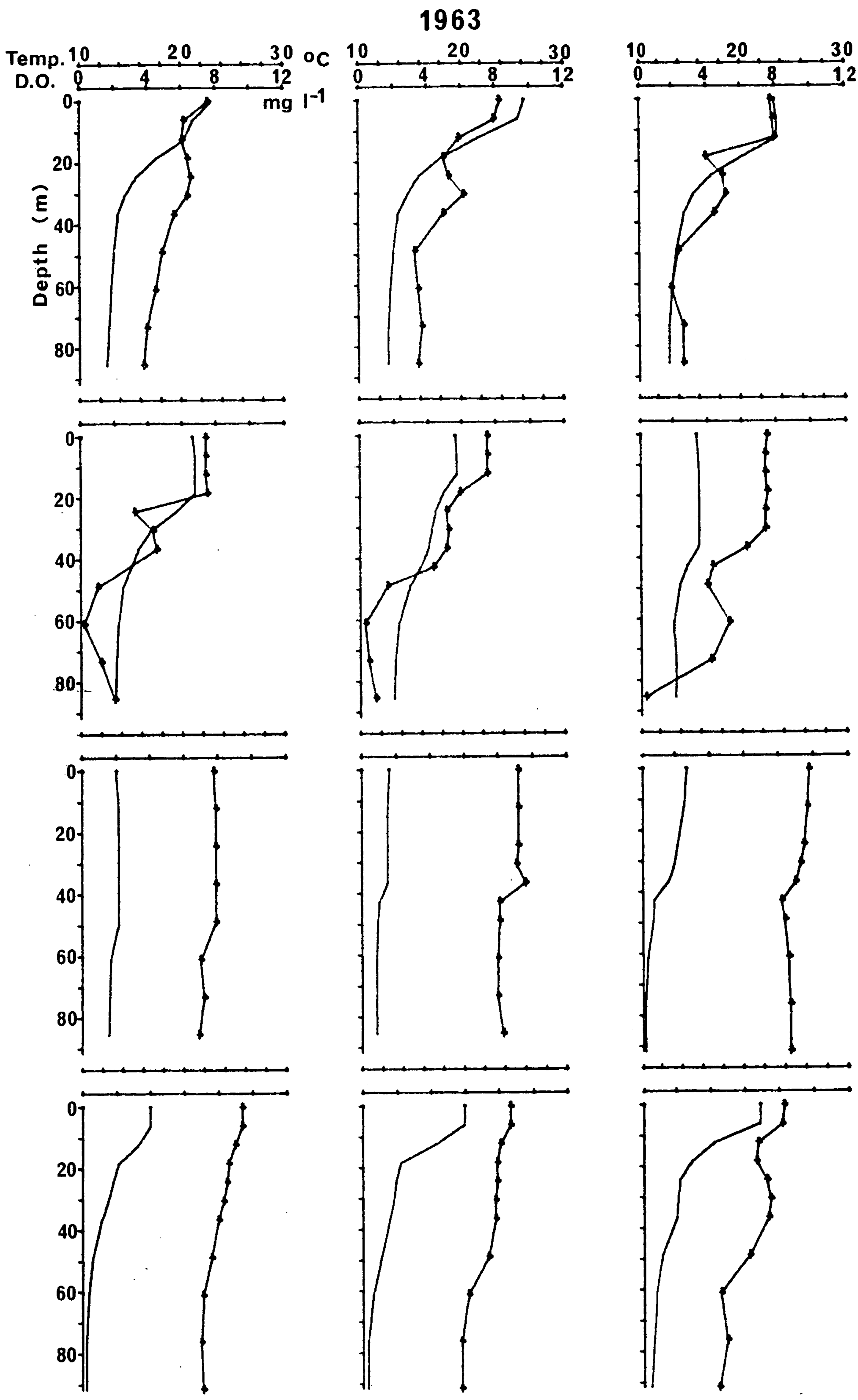
JAN. FEB. MAR.
APR. MAY. JUN.
JUL. AUG. SEP.
OCT. NOV. DEC.

Axes are of equivalent scale:-

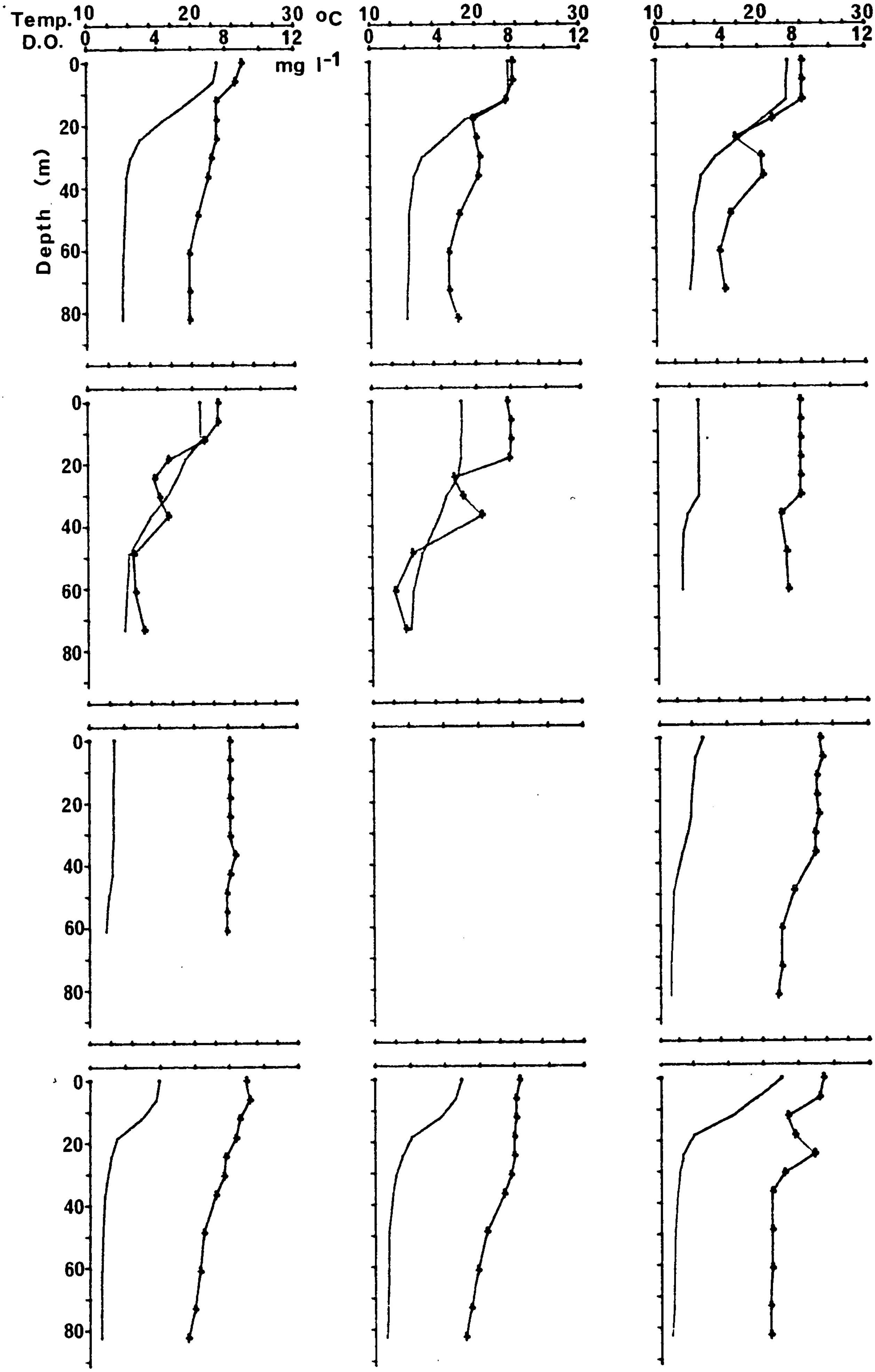
Temperature; 10 - 30°C, with 2°C increments
Oxygen; 0 - 12 mg l^{-1}, with 2 mg l^{-1} increments
Depth; 0 - 91.4 metres, with 10 m increments

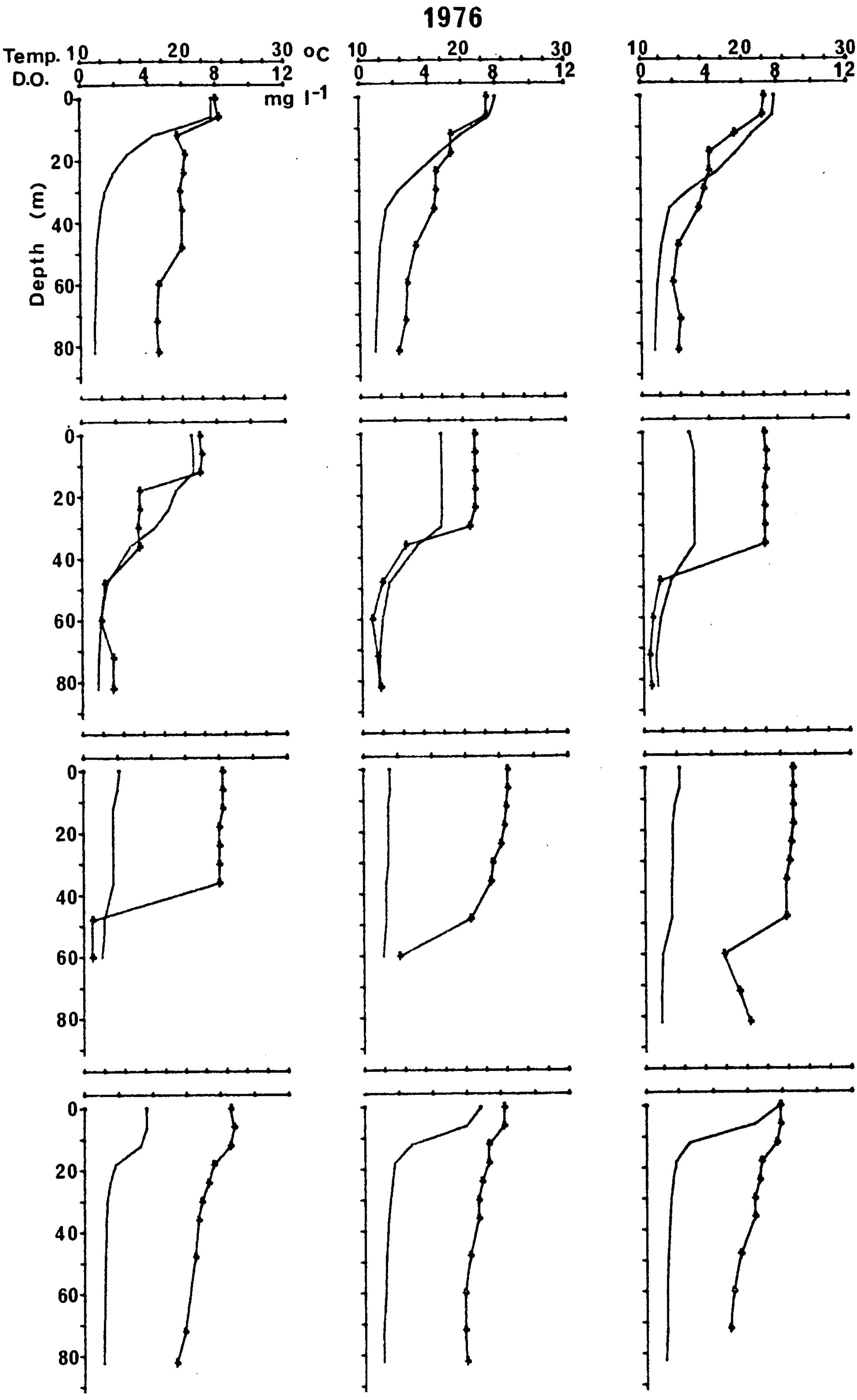
Symbols:-

Temperature (·)
Dissolved oxygen (+)


1963
Temp. 10 20 30 oC
D.O. 0 4 8 12
mg l-1
Depth (m)
0
20
40
60
80

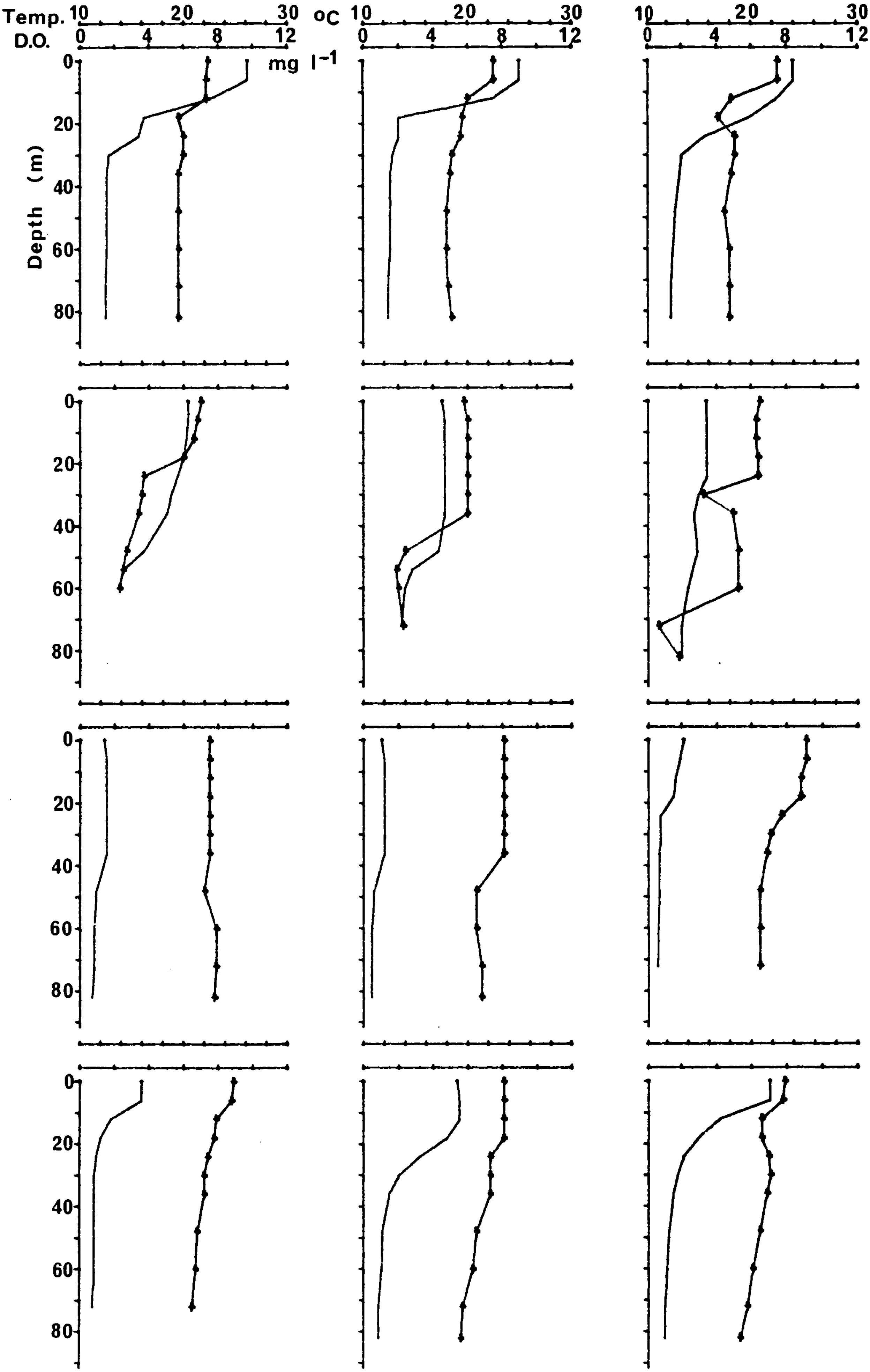
1974

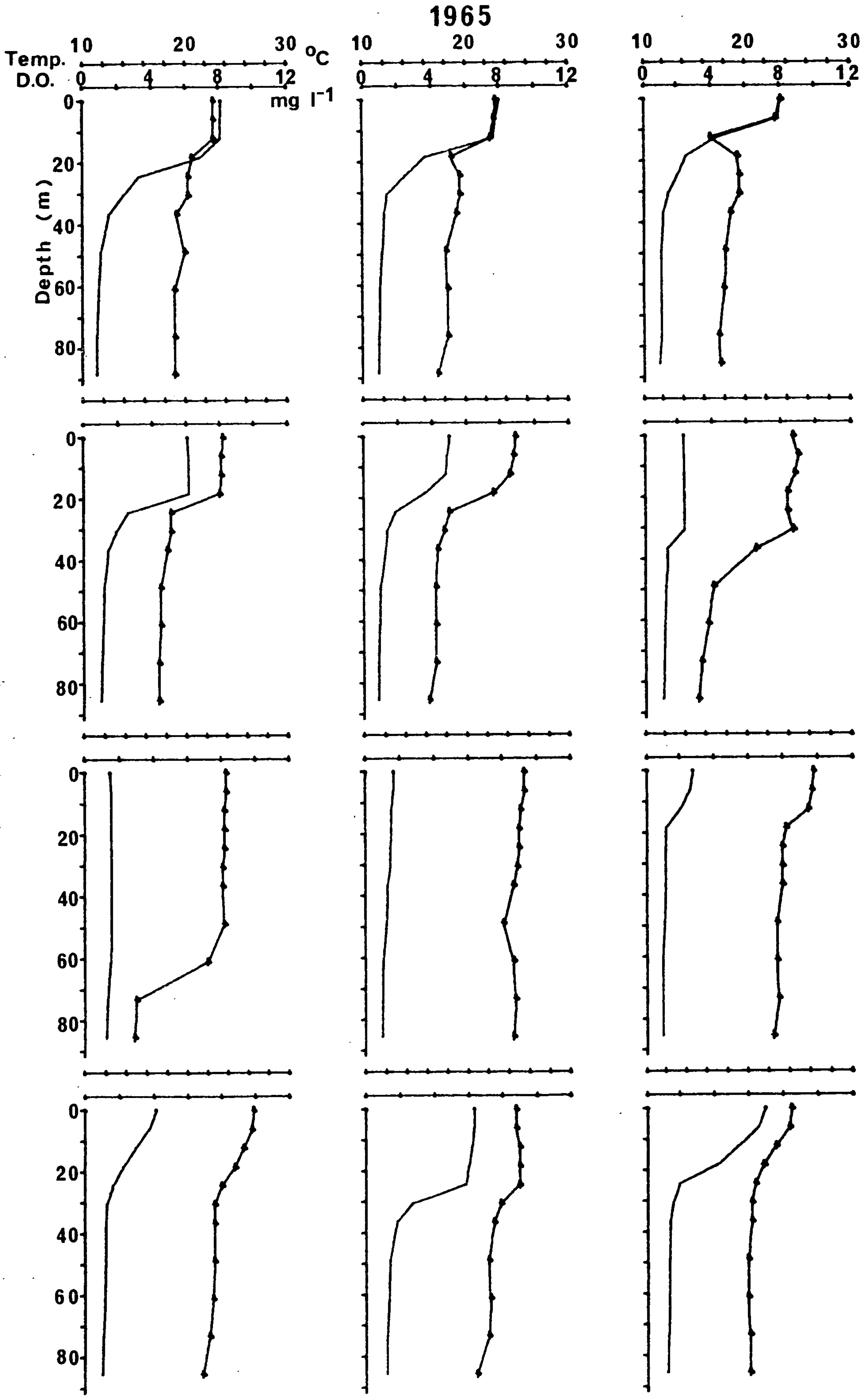


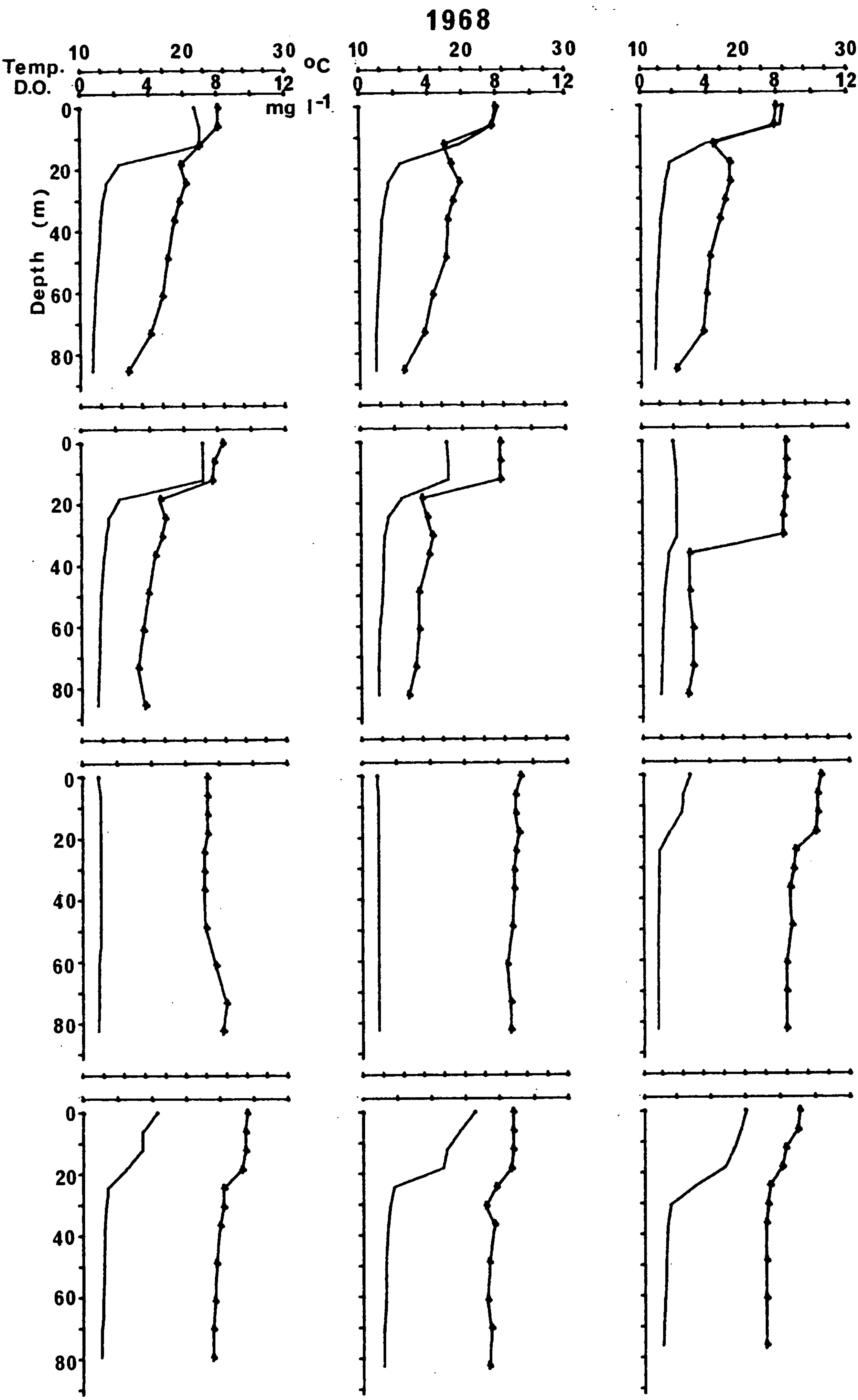


1976
Temp. 10 20 30 oC
D.O. 0 4 8 12 mg l-1
Depth (m)
0
20
40
60
80

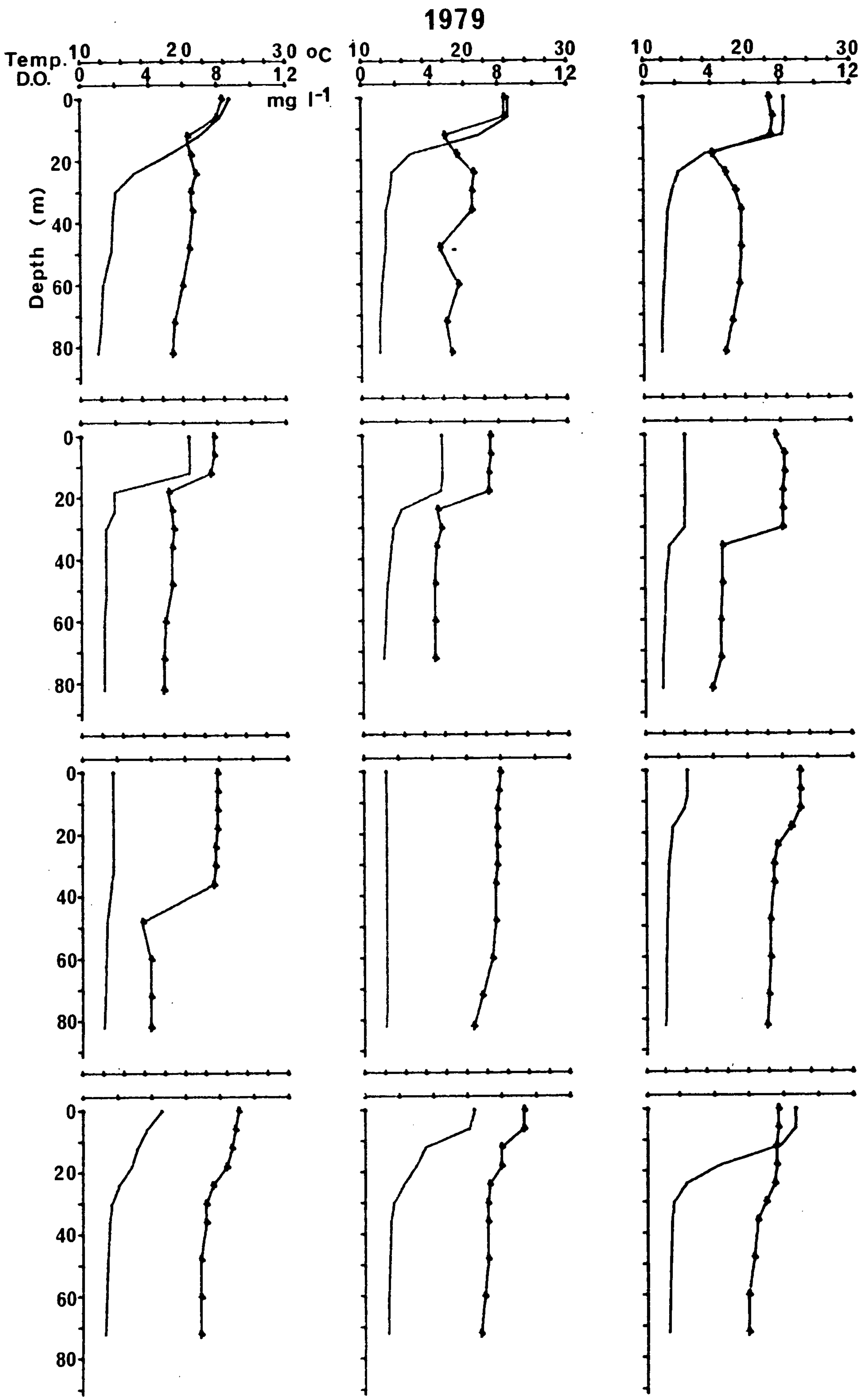
1978

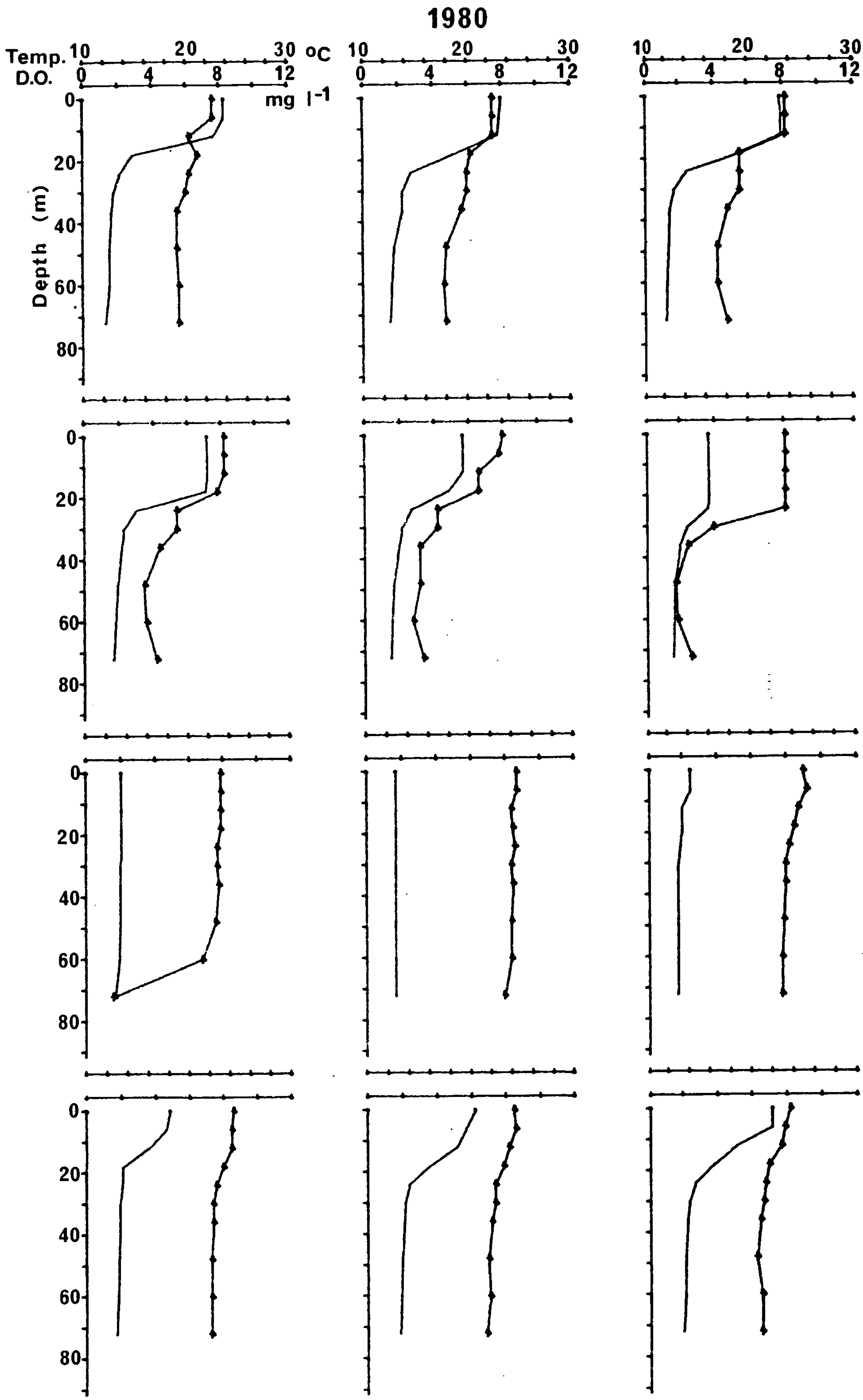



1965
Temp.
D.O.
10 20 30 °C
0 4 8 12
mg l⁻¹
Depth (m)
0 20 40 60 80



1968
Temp.
D.O.
10
20
30
oC
0
4
8
12
mg l-1
Depth (m)
0
20
40
60
80


1979
Temp. 10 20 30 oC
D.O. 0 4 8 12
mg l-1
Depth (m)
0
20
40
60
80


1980
Temp. 10 20 30 oC
D.O. 0 4 8 12
mg l-1
Depth (m)
0 20 40 60 80

APPENDIX 3

This appendix contains the text of a paper:-

Ferris, J. M., and Burton, H. R. (1985). The annual cycle of heat content and mechanical stability of hypersaline Deep Lake, Vestfold Hills, Antarctica.

which has been submitted to Hydrobiologia, as part of a planned volume containing the proceedings of a conference on research in the Vestfold Hills (Antarctica), held in Hobart in August 1984. The paper is included here primarily as evidence of other work completed by me during my post-graduate studies, but has in common with the present work the application of the early concepts of heat budget, stability, and the work of the wind to the understanding of stratification processes in lakes.

The Annual Cycle of Heat Content and Mechanical Stability of Hypersaline Deep Lake, Vestfold Hills, Antartica.

J.M. Ferris and H.R. Burton

keywords: heat budget, stability, saline lake, Antartica.

ABSTRACT

Deep Lake (Vestfold Hills) is a hypersaline lake (c. 10 x seawater concentration), with approximately seawater ionic proportions, which rarely freezes and is characterised by a monomictic thermal cycle. Winter circulation, at c. -17° C, lasts for 2-3 months. In summer epilimnetic temperatures from 7-11° C result in large vertical thermal gradients (21-26° C). This combines with the enhanced rate of density change per degree Centigrade that accompanies such high salt concentration, to produce a particularly stable density configuration in Deep Lake (Schmidt stability c. 8000 gm-cm cm^{-2}). The Birgean annual heat budget (c. 24500 cal cm^{-2}) is comparable to that of a temperate lake with a similar mean depth, despite the particularly inefficient heating indicated by a high ratio of Birgean wind work to annual heat budget (0.37 gm-cm cal^{-1}). Deep lake retains c. 50% of the incident solar radiation during the short summer heating period, a similar efficiency to "first class" lakes in North America. Extended daylight hours undoubtedly contribute to the high maximum rate of heating in the lake (444 cal cm^{-2} day^{-1}). Deep Lake cools at a rate less than half of its average heating rate. Although the lake's stability is dominated by thermal gradients, partitioning of the total stability into thermal and saline components shows that salinity can contribute up to c. 20% of the maximum summer Schmidt stability. Salinity gradients have the following seasonal characteristics. In early summer the effect of a small melt-stream is to increase stability by diluting the epilimnion. In autumn evaporative water loss can overtake this effect, creating small de-stabilising salinity gradients. The usually short-term stabilising influence of snow-fall and drift is less predictable, but is probably most common in winter when strong wind events are more frequent. Hypersalnity has a profound effect on the physical cycle of Deep

Lake, through freezing point depression and the increased rate of density change with temperature. These changes affect the lake's biota, both in relation to osmotic stress, and by effectively exposing them to a more thermally extreme environment. A comparison between Deep Lake and a smaller lake of similar salinity (Lake Hunazoko, Skarvs Nes), demonstrates that it is inappropriate to consider the biological effects of salinity in isolation. The smaller lake offers warmer epilimnetic conditions for at least part of the summer, which may explain the much greater limnetic algal production in Lake Hunazoko.

INTRODUCTION

The study of empirical parameters, such as heat budget and mechanical stability, which can be derived from thermal profiles and a knowledge of lake morphometry, is uncommon for Antartic lakes. A study of the annual cycle of heat content and stability of an Antartic saline lake is rarer still.

Deep Lake (Vestfold Hills) has a number of features that make it of particular importance in the context of such studies. The lake is hypersaline (c. 10 x SW; Barker 1981), deep (36m), and very clear, having the rare distinction of remaining ice-free in most years. In this last fact is the suggestion that Deep Lake may provide a record of the outer limits of thermal behaviour in a climate distinguished by its extremes. Further, Deep Lake may give a more general insight into the role of termperature and dissolved salts in the annual stratification cycle of saline lakes.

MATERIALS AND METHODS

Thermal profiles have been intermittently recorded, for Deep Lake, since 1957 (Kerry et al 1977). Since 1973 there have been two periods of regular sampling, October 1973 to January 1976, and January 1977 to March 1979. This latter period of relatively intensive measurement is considered in detail here. Temperature measurements were taken using a thermistor designed to give probe/battery voltage ratio, to overcome cold effects. Profiles, with a depth interval of 1 m were taken, in the deepest section of the lake, weekly during 1978 and early 1979. Sampling was less frequent, both temporally and spatially in 1977. Fig. 1 shows the sampling distribution for the study period. Water samples were collected with a 2.5 l Van Dorn bottle, from sufficient depths to characterise vertical salinity changes for each of the sample dates.

These samples were stored in tightly sealed heavy polyethylene bottles which were returned to Australia where density (20° C) was measured using an Anton Parr DMA55 density meter, capable of reading reproducibly to 5 decimal places. These density measurements were used to estimate salinity (g kg^{-1}) from the relationship published by Witfield & Jagner 1981. The morphometry of Deep Lake was derived from echo-sounding transects made in December 1978.

The Idso (1973) modification of Schmidt stability, Birgean wind-work, heat content, and volume weighted averages of temperature and salinity were calculated using a version of LIMNO (cf. Johnson et al 1978) modified to accept direct input of denisty profiles, rather than inferring density from temperature and salinity. In situ density was estimated from measured density (20° C) using an experimentally determined linear temperature/ density relationship for Deep Lake, prior to use of LIMNO. Subsequently a general algorithm (DEQCLC), relating temperature and density for saline lakes in the Vestfold Hills, ranging from, approximately seawater salt concentration to that of Deep Lake (c. 10 x SW), has been completed. DEQCLC is based on a series of quadratic regressions, and was used to provide data presented in Fig. 3. Stability was partitioned into thermal and saline components by first calculating total stability (S_{tot}), and then declaring the water isohaline at the volume weighted mean salinity, to enable a separate calculation of thermal stability (S_t). Saline stability (S_s) then equals S_{tot} - S_t.

THEORETICAL CONSIDERATIONS

Heat budget, stability and wind-work are empirical quantities derived from thermal and salinity profiles, and a knowledge of lake morphometry.

The Annual heat budget of a lake is the term used to denote the change in stored heat between the time of minimum heat content and the subsequent time of maximum heat content. It is the amount of heat stored by the lake during the annual heating period, expressed per unit surface area of the lake. In the southern hemisphere the sequential nature of Birge's heat budget often means that it is calculated as the difference between the winter minimum of one calendar year and the maximum occurring early in the following year. Hypersalinity, in Deep Lake, introduces a potential complication in calculation of the heat content, as stored heat is a function of mass, temperature and specific heat. In a freshwater system mass and specific heat are unity, so that the heat content (calories) is calculated simply as temperature x volume. In hypersaline water, however, both mass and specific heat are significantly altered. Fortunately it appears that increased mass (1.176 gm cm^{-3}) is approximately balanced by lowered specific heat (c. 0.82 cal gram-$°C^{-1}$), and we have assumed that temperature x volume adequately reflects heat content in Deep Lake. We believe that any error in this assumption is less that 5%. Mason (1967) adopted the same approximation for calculating the heat content of Mono Lake, California (c. 2 x SW).

Birge's work of the wind (B), and Schmidt's stability (S) are complementary quantities expressing work done through the lake surface, firstly to establish a given density stratification, from an assumed initial condition, and then to destroy that stratification, without further exchange of heat or solute. References, which deal with the mathematical and conceptual framework of these quantities in more detail than is given here, include Birge (1916), Hutchinson (1957), Idso (1973), Walker (1974), Johnson et al (1978), and Viner (1984).

Birgean wind work (B) is limited by its conceptual time dependence, in that a starting point is assumed, several months prior to the period of maximum stratification in summer. Simplified assumptions are necessary concerning the processes which led to the final maximum departure from the initial condition. In the context of Deep Lake, the likely contribution of salinity to this density stratification complicates the interpretation of B. In general, processes that contribute to density layering but do not operate through the surface of the lake, for example inflows of different temperature, turbidity or salinity, fall outside the conceptual framework of Birgean wind work.

Schmidt Stability, on the other hand, is conceptually instantaneous, and is therefore not limited by assumptions concerning the establishment of the existing density stratifications. It does however assume that the means of destroying the stratification is turbulent energy, transferred to the lake through its surface (ie by the wind). Finally it should be noted that all these quantities are minima. That is, the Heat budget is the minimum quantity of heat stored in the lake during the period of heat income. Birgean Work and Schmidt Stability are minimum estimates of energy transferred to a lake through its surface, by the wind.

RESULTS AND DISCUSSION

Thermal Stratification

Descriptive reports of thermal stratification in Deep Lake, based on data collected between 1973 and 1976, are contained in Kerry et al (1977), Campbell (1978), Burton and Campbell (1980), and Barker (1981). Temperature isopleths for the period 1977-1979 are shown in Fig. 2. In the absence of ice-cover, Deep Lake exhibits an essentially monomictic thermal cycle. Vertical circulation is normally complete for a period of 2-3 months between July and September (winter), when water temperature falls to between -17 and -18° C. Maximum surface temperatures range from 7-11.5° C for brief periods in summer. Bottom temperatures rarely warm to -14° C (except for a small pocket below 34m) giving a total range of 4-5° C annually. The maximum vertical temperature gradients in summer, from 21-26° C, are large in relation to most other lakes. A further point of interest is that temperatures exceeding 0° C are only recorded in the upper 0-10m of the water column, and for a period of 2-3 months of each year.

Barker (1981) noted the existence of a deep layer of more saline water, that appeared to survive the 1974 mixing period (which extended only to between 25 and 30m), but was obliterated in the winter mixing of 1975. The extension of sampling to 36m, in 1978, revealed a small pocket of more saline water which supported an inverse temperature gradient. This has been found to exist on occasional visits to the lake since that time. It is possible that a small pool, of variable vertical extent, exists for much of the time in Deep Lake. It is most probably generated by dissolution of crystalline bottom deposits of mirabilite ($NA_2SO_4 . 10H_2O$) which occur in the lake.

<u>Annual Terms</u>

Table 1 contains a list of mainly high latitude lakes for which some data, concerning annual heat budget, or annual maximum, of either B or S, is published.

Deep Lake has a large annual heat budget compared to these lakes, except for much larger Great Slave Lake, and Lake Chandler if the latent heat of fusion is taken into account. Goreham (1964) provides predictive regressions relating lake morphometry to the magnitude of the annual heat budget for 71 temperate lakes. In this context Deep Lake's annual heat budget closely matches that predicted for lakes of similar mean depth, but is considerably larger (c. 9000 cal cm^{-2}) than would be predicted on the basis of either surface area or volume.

The difficulty of interpreting Birge's Wind Work for a lake where salinity can play a significant role in determining the density stratification has already been mentioned. Nevertheless, as the most important salinity effects in Deep Lake relate to dilution of the surface water, it is apparent that work is done in mixing this lighter water downwards in a way analagous to that for water warmed by the sun. Consequently the estimate of Birgean Work, in Table 1, incorporates the influence of both temperature and salinity on the change from the previous period of minimum stratification to the subsequent time of maximum deviation from this initial condition. Calculation of B for polar lakes is rare, presumably because most such lakes are ice-covered for the greater part of the year. However, the B value for Deep Lake (9000 gm-cm cm^{-2}) is unusually large compared to that of hypersaline Mono Lake, and also in the broader context of temperate and tropical lakes throughout the world. Large Freshwater lakes, such as Lake Rinihue (39° 50' S : Campos et al 1978), and Lake Powell (37° 30' N : Johnson and Merritt, 1979) yield values of B no greater than ~~6~~8000 gm-cm cm^{-2}.

Hutchinson (1957) used the ratio of B to summer heat income (equivalent to heat budget for monomictic lakes) to give an efficiency term (gm-cm cal^{-1}) describing the work required to account for each calorie of heat stored by the lake; a low ratio indicates efficient heating, such as for Lake Chandler (Table 1), while the higher ratio for Mono Lake suggests relatively inefficient heating. Data for ten temperate and tropical lakes range from 0.06 (Pang-gong Tso ; 33° 45' N, high altitude; Hutchinson, 1957) to 0.56 (L. Valencia; 10° N, low altitude; from data given by Lewis, 1984). Although Deep Lake (0.37) lies within this range the efficiency ratio is second only to that of Lake Valencia, indicating particularly inefficient heating relative to most lakes for which data is available.

Deep Lake is characterised by a relatively high stability compared to lakes represented in Table 1, with an annual maximum S of similar magnitude to that reported by Hutchinson (1957) for much larger and deeper Pyramid Lake.

Density Change with Temperature

A principle factor contributing to the high stability and inefficient heating of Deep Lake is the very high concentration of dissolved salts. Mason (1967) observed for Mono Lake (c. 2 x SW) ": a very rapid change of density with temperature .". Deep Lake has a concentration approximately 10 x SW (Barker 1981), and shows a much greater density change per degree Centigrade. The rate of density change with temperature (0-20° C) is illustrated, for freshwater, seawater, and Deep Lake water, by the slopes of the curves in Fig. 3. Direct comparison of equivalent volumes of fresh and Deep Lake water warmed from 4° C to 9° C (within the range of temperatures recorded for Deep Lake) shows that the change in density is approximately 11 times greater for Deep Lake water. The density change per degree Centigrade, of freshwater, increases with temperature above 4° C, in the range of temperatures normally encountered in lakes (Hutchinson, 1957). This change in density of Deep Lake water (between 4° C and 9° C) is comparable (c. 1.3 times) to that of freshwater warmed from 30° C to 35° C. Therefore, Deep Lake will tend to stratify as a result of quite small thermal gradients, in much the same way as a tropical Freshwater lake. The generally low stability of tropical lakes reflects the very small thermal gradients forming under conditions of diminished seasonality. In Deep Lake large summer thermal gradients combine with rapid density change per degree Centigrade to give a quite stable density configuration explaining both high S values and inefficient heating. The thermal density gradient recorded on 14.1.77 was approximately twice that of a freshwater lake stratified from 35° C to 4° C.

Annual Cycles

Figures 4-5 show the annual progression of heat content and stability in Deep Lake. The stability has been partitioned such that total and thermal stability are shown, the gap between the two lines represents the stability resulting from saline gradients in the water column. The apparent smoothness of the 1977 curves is a result of less frequent sampling in that year.

Heat content, and therefore volume weighted average temperature, never exceed zero. Heat content shows an almost saw-toothed progression characterised by near linear heating and cooling phases with rapid interchanges between them. It is also obvious that heating proceeds more quickly but for a shorter time than the cooling phase. Table 2. contains comparative data on heating and cooling rates from various lakes around the world, most of which are freshwater lakes. These rates are simply change in heat content per unit time for periods of approximately linear gain or loss of heat (cal cm^{-2} day^{-1}). Deep Lake has a relatively rapid period of heat gain from September to December, though it is less than for Lakes Powell and Baikal. The maximum rate, however, is very large, and this most probably reflects the very long hours of daylight in summer.

Deep Lake cools at a rate less than half of its heating rate, and is slow to cool compared to other lakes (Table 2). The maximum rate of heat loss, however, is relatively large. In his considerations of lake thermal capacity, Hutchinson (1957) estimates the percent retention of incident radiation by temperate lakes. He noted that 40-60% retention was characteristic of "supposedly first class lakes" in North America, though 70% or more was possible for large lakes such as Baikal and Michigan. Using data for solar radiation for nearby (5km) Davis station (Burton and Campbell, 1980) it is estimated that Deep Lake retained 50% of incident solar radiation in the period September - December 1977. This is despite the inefficiency of heating in the lake, its small size, and the comparatively low angle of direct solar radiation. Mason (1967) estimated that Mono lake retained only 22% of incident radiation during the heating period from March - August. Climatic differences may be invoked to explain this latter comparison, as the mean wind speed at Davis (5m sec^{-1}) is about twice that for Mono lake. Further, records of winds greater than 15m sec^{-1} are rare for Mono lake, while winter winds at Davis frequently exceed this figure.

It is evident, from Fig. 5, that temperature is the dominant contributor to S_{tot}, but that salinity is significant at times. In winter, when S_t is minimal, salinity gradients can be the dominant term in S_{tot}. Total stability shows a three part annual progression. A rapid, and approximately linear decline in stability (January - April) follows the maximum of S_{tot} in January. Between April and about July the decline in S_{tot} is slower. July - September contains the period of minimum winter stability. The ensuing heating period (October - December) results in a near linear increase in S_{tot}, at a rate greater than for any other part of the year. The assymetry of this annual cycle of stability is in accord with that suggested by a statistical treatment of data for eleven New Zealand lakes by Viner (1984).

Regardless of different sampling frequencies, it is apparent, from Fig. 5, that the contribution of salinity to S_{tot} was qualitatively different in 1977, compared to 1978. In 1977 S_{tot} had a consistent component due to salinity, except for the period of winter mixing, whereas in 1978 the influence of salinity was irregular and episodic.

In Fig. 6 saline stability (S_s) is presented on an enlarged scale. Fig. 7 shows the variation in surface and 34m salinity concentrations. The absolute magnitude of stability resulting from salinity gradients is evident from Fig. 6. The maximum effect, 1470 gm-cm cm^{-2} occurred in January 1979 when the saline stability was approximately 20% of the maximum stability for that summer. Direct comparison of Fig. 6 and Fig. 7 gives clear indication that fluctuations in surface salinity generate the changes in S_s. These two graphs are, in essence, mirror images, with the sudden spikes of stability reflecting equally short-lived changes in surface water salinity. Despite this irregular progression of saline stability there are consistent features. The data suggests, that a salinity gradient is consistently found in the summer, particularly December - January. This probably results from the single melt-stream which drains into Deep Lake for 1-2 months in summer (Barker 1981). This salinity gradient declines between January and April, partly as the superficial fresher layer is mixed into the lake, and partly as a result of evaporative concentration during this period of reduced wind, and therefore drifting snow (Burton and Campbell 1980). This latter effect causes the slightly negative S_S recorded between February and April of both 1978 and 1979.

The slightly elevated surface salinity, which accompanies these de-stabilizing salinity gradients, is shown in Fig. 7. A maximum negative stability of -210 gm-cm cm^{-2} was recorded in April 1978. Dilution of the surface water by direct snowfall, or snowdrift during strong-wind events, can occur at any time of the year. Burton and Campbell (1980) suggest that Deep Lake acts as a snow drift-trap, in which case the input of drift snow will be greatest in winter when strong-wind events are most common. This probably explains the sudden increases in saline stability in October 1977 and the "spikes" in July, September and October of 1978. This stabilising effect of precipitation or inflow is likely to characterise saline lakes in general. In Deep Lake this gives rise to the paradox that periods of strong winds do not simply promote vertical mixing, as in freshwater lakes, but also tend to dampen mixing by contributing a superficial layer of less dense water to the lake.

Profiles of Temperature and Salinity

Fig. 8 contains representative profiles of temperature and salinity for each month of the study period . These illustrate the general dependence of salinity gradients on the thermal structure in Deep Lake.The profiles clearly show that salinity changes are surface phenomena whose vertical extent usually coincides with that of the epilimnion. The early part of 1977 appears to be an exception to this pattern, as the main halocline lies well below the epilimnion. Closer examination, however, reveals that the halocline coincides with a secondary, deeper, thermocline. The formation of secondary thermoclines appears to be a common phenomenon in Deep Lake, during the heating perod (eg December 1978), and presumably results from interplay of calm and windy weather at a time when thermal stratification is being re-established. The mixed depth of a lake is subject to considerable variability, especially during the early stages of the heating period (Darbyshire and Edwards 1972; Mortimer 1974; Wetzel 1983). In a hypersaline lake the stabilising effect of precipitation will certainly play some part in determining the mixed depth as well.

The effect of snow fall or drift on the lake is best illustrated by the October 1978 profile which shows a very thin layer of relatively less saline water. Fig. 8 also shows the effect of evaporation on the epilimnetic salinity; slightly higher salt concentrations are evident in February and March of 1979.

Deep Lake and Lake Hunazoko

Deep Lake (Vestfold Hills) and Lake Hunazoko (Skarvs Nes) are lakes of similar salinity, but markedly different limnetic algal production. Selected comparative data for the two lakes are presented in Table 3.

Despite some disagreement over the species diversity in these lakes, a problem common to the investigation of hypersaline lakes throughout the world (see Nissenbaum 1979; Heywood 1984), it is accepted that both lakes support active populations of the halophilic green alga Dunaliella sp. Further, it is evident that Lake Hunazoko supports limnetic populations of Dunaliella reaching biomass concentrations (chlorophyll-a) that exceed those of Deep

Lake by about an order of magnitude (cf. Akiyama 1975; Tominaga & Fukui 1981; Wright & Burton 1981; Heywood 1984). The extremely low limnetic production in Deep Lake has been discussed by several authors (cf. Campbell 1978; Williams 1979; Hand 1980; Wright & Burton 1981), and various factors have been proposed as growth limiting in the lake. these include nutrient availability, salinity, and temperature.

Concentrations of nitrogen and phosphorus tend to be higher in Deep Lake than in Lake Hunazoko (Table 3), but Campbell (1978) and Williams (1979) have suggested that, in situ, nutrients may be unavailable in Deep Lake. The absence of detectable ammonia, probably the preferred form of nitrogen for planktonic assimilation (Brezonik 1972), in Deep Lake, compared to levels of 24 - 84 $\mu g\ l^{-1}$ in Lake Hunazoko (Tominaga & Fukui 1981), coupled with the low ratio of inorganic-N to inorganic-P in Deep Lake, compared to Antarctic Ocean water (cf. Burton 1981) support the possibility of nitrogen limitation, without in any way establishing it's existence in Deep Lake.

Wright & Burton (1981) suggest that the existence of a lake of similar salinity, but with much greater limnetic algal production (L. Hunazoko) shows that "... salinity alone is not limiting productivity in Deep Lake.". However, there are salinity differences between the two lakes, in that Lake Hunazoko has a much more variable surface salinity, and only reaches salinities within the range found in Deep Lake, in its hypolimnion. This suggests that summer inflow of meltwater constitutes a greater percentage of the total lake volume in Lake Hunazoko, which could have a variety of ecological effects in the lake.

The most obvious differences between the two lakes, however, are that Lake Hunazoko is smaller, shallower, and more turbid. Thermal profiles, taken two days apart in January 1977, show that Lake Hunazoko had a shallower and warmer epilimnion (Table 3), and Watanuki & Ohno (1975) report a surface temperature some 3° C greater than any recorded for Deep Lake in the period

1973-79. How long this situation persists, in any year, cannot be determined from the available data. Lake Hunazoko might be expected to heat and cool more rapidly than Deep Lake, but during the cooling phase, when epilimnetic temperature is determined by the progressive incorporation of cold hypolimnetic water as well as by surface heat exchange, it is possible that epilimnetic temperature would decline most rapidly in Deep Lake, where a much greater percentage of the total lake volume remains cold throughout the year.

In general, therefore, the effects of salinity are not independent of other physical factors. The effect of increasing lake size and volume in Antarctic hypersaline lakes will probably be to restrict the summer epilimnial temperature, and to shorten the period in which relatively warm conditions are available for algal growth. Parallel investigation of these lakes, for at least one annual cycle, would be valuable in defining the interaction of factors which limit algal growth in Antarctic hypersaline lakes.

SUMMARY AND CONCLUSIONS

Deep Lake is the most ionically concentrated lake of the Vestfold Hills, with relative ionic proportions similar to sea water. As a direct result of its hypersalinity the freezing point of Deep Lake water is depressed to the point where the lake virtually never freezes, despite water temperatures of -18° C, during winter circulation. A second consequence of the lakes high salt content is an increase in the slope of the temperature/density relationship. In combination with particularly high summer thermal gradients Deep Lake therefore exhibits a very stable density configuration. A further effect of this enhanced rate of density change with temperature is that the lake has very inefficient heating, as indicated by the ratio of Birgean work to summer heat income. Nevertheless, Deep Lake has an annual heat budget close to that expected of temperate lakes with similar mean depth, and captures c. 50% of the radiation incident upon it during the short heating period, a similar efficiency to those of "first-class" lakes in North America. It is suggested that water clarity and relatively frequent strong winds counter-balance the density and viscosity factors tending to retard downward mixing in the lake. The maximum rate of heating (444 cal cm^{-2} day^{-1}) inferred from changing heat content with time, are comparable to the greatest anywhere in the world, despite the inefficiency of heating and low solar angle. Extended summer daylight hours are undoubtedly important in this. Deep Lake cools at a rate less than half of its average heating rate.

The annual cycle of mechanical stability is a function primarily of thermal stratification, which sets this lake apart from the many meromictic lakes of this region. Salinity gradients are capable of significant contribution to total stability, and have the following seasonal characteristics. During early summer the effect of a small melt-stream is to increase stability by diluting epilimnial water. Later in summer and in autumn the evaporative loss from the lake can reverse this dilution effect creating small de-stabilising salinity gradients. Salinity gradients do support inverse thermal gradients in Deep Lake. These are mostly transient features of the cooling phase, but a small pool (below 34 m) of markedly greater salinity was found in 1978, and may be persistent. Surface dilution by falling and drifting snow is a less predictable salinity effect which usually generates short-term increases in saline stability. These events may be more common in winter when strong winds, and hence drifting snow, are most frequent (Burton & Campbell 1980). If surface dilution occurs during the period of thermocline re-formation, then relict thermoclines with associated haloclines may persist until the following cooling phase (eg. 1977).

It is evident that hypersalinity has a profound effect on the physical cycle of a lake, particularly a polar lake, by depressing the freezing point and increasing the rate of density change per degree Centigrade, compared to freshwater lakes. These changes affect the biota of the lake also, both in relation to osmotic stress, and by effectively exposing them to a more thermally extreme environment. Deep Lake is characterised by a very brief period when epilimnetic temperatures exceed 0° C, in any year. A comparison between Deep Lake and a smaller lake of similar salinity (Lake Hunazoko, Skarvs Nes), demonstrates that it is inappropriate to consider the biological effects of salinity in isolation. The smaller lake offers warmer epilimnetic conditions for at least part of the summer, which may explain the much greater limnetic algal production in Lake Hunazoko. However, further investigation of both lakes is needed to determine the relative significance of thermal regime and other potentially limiting factors, such as nitrogen availability in Deep Lake.

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Table 1 Comparative data for annual heat budget, Birgean work, and mechanical stability, for mainly high latitude lakes. Ordered, where possible by mean depth.

Lake	Latitude	Altitude (m. ASL)	Area (km^2)	Depth Maximum (m)	Depth Mean (m)	Annual Heat Budget (cal cm^{-2})	Birgean Wind Work (gm-cm cm^{-2})	Heating Efficiency (gm-cm cal^{-1})	Schmidt Stability (gm-cm cm^{-2})	Source
Deep	68° 34' S	-50	0.64	36	20.2	24600	c. 9000	0.37	8100	
Tennesvatn	67° 53' N	230	1.0	168	93.0	5000 [a]				Hutchinson (1957)
Great Slave	61° 31' N	150	27200	614	62.0	24300 +				Hutchinson (1957)
Schrader	Alaska		13.2	57	33.0	8980				Wetzel (1983)
Mono	38° 00' N	1979	199.5	51.5	19.0	22300	5750	0.26	3500	Mason (1967)
Chandler	67° 40' N	886	15.0	21	13.5	5760 [a,b]	175	0.03		Hutchinson (1957)
Burton	68° 34' S	0	1.35	18.3	7.2				1520	Burton (unpublished)
Irizar	62° 58' S	7	0.046	18	4.3	1430			770	Bienati (1967)
Joyce	c. 77° 30' S								2610	Parker & Simmons (1978)
Fryxell	77° 30' S								1140	Parker & Simmons (1978)
Chad	77° 39' S								550	Parker & Simmons (1978)

[a] Summer heat income, less than the annual heat budget for a dimictic lake.

[b] Heat budget = 27000 cal cm^{-2}, if the energy needed to melt the ice cover is included.

Table 2 Heating and cooling rates for various lakes, (Change in heat content per unit time).

Lake	Heating Period	Heating Rate Mean ($cal\ cm^{-2}\ day^{-1}$)	Heating Rate Maximum	Cooling Period	Cooling Rate Mean ($cal\ cm^{-2}\ day^{-1}$)	Cooling Rate Maximum	Source
Deep (Antarctica)	Sep. - Dec.	252	444	Jan. - Sep.	-107	-339	
Powell (USA)	Apr. - Aug.	368	439	Sep. - Jan.	-345	-392	Johnson & Merritt (1979)
Kamloops (Br. Col.)	June		c. 400	January		c. -400	Carmack et al (1979)
Baikal (USSR)	117 days	362					Hutchinson (1957)
Klammingen (Sweden)	May		332	October		-201	Hutchinson (1957)
Riñihue (S. Am.) [a]	December		300	July		-260	Campos et al (1978)
Green (USA)	May - Aug.	215					Hutchinson (1957)
Mono (USA)	May - Aug.	176					Mason (1967)
Ladoga (USSR)		160					Hutchinson (1957)
Windemere (Gr. Br.)		160					Hutchinson (1957)
Titicaca (S. Am.)	Sep. - Dec.	158		May - Jul.	-172		Richerson et al (1975)
Burragorang (Aust.)	Sep. - Dec.	113	125	Apr. - Jul.	-157	-169	Ferris (unpublished)
Neagh (Ireland) [b]	May		106	November		-106	Gibson & Stewart (1973)
Augher (Ireland)	Apr. - May		76				Rippey (1983)

[a] These rates apply to the surface layer only.

[b] Predictions from an empirical model. content per unit time).

Table 3 Comparative data for Deep Lake and Lake Hunazoko. Thermal data for Deep Lake are from the present study, and the remainder from Barker (1981). Unless otherwise indicated, data for Lake Hunazoko are from Tominaga & Fukui (1981).

Parameter	(units)	Deep Lake	Lake Hunazoko
Surface area	km^2	0.64	0.142
Maximum depth	m	36.0	9.2
Mean depth	m	20.2	c. 4 - 5
Ice cover		none	probably none
Temperature	°C	7.5 to -15.7 (26.1.77)	12.0 to -13.0 (24.1.77)
Minimum temp.	°C	-17.5	-18.3
Depth of thermocline	m	9 - 18	3 - 4
D.O.	$ml\ l^{-1}$	2.1 - 2.6	2.6 - 2.8
Chloride	$g\ l^{-1}$	154.8 - 168.3	73.1 - 158.0 [a]
pH		7.4 - 7.5	7.3 - 7.7 [a]
Euphotic depth	m	> 36	3
Phosphate	$\mu g\ l^{-1}$	56.7	10.9
Nitrate	$\mu g\ l^{-1}$	82.2	20.3

[a] Data also taken from Watanuki & Ohno (1975).

FIGURE CAPTIONS

Fig. 1 A Dot-matrix of the sample points, plotted against depth and time, which were used to construct the temperature isopleth diagram.

Fig. 2 Temperature isopleths (°C) for Deep Lake, from January 1977 to March 1979.

Fig. 3 A comparison of density (gm cm^{-3}) / temperature (°C) curves for freshwater (upper panel), seawater (middle panel), and Deep Lake water (lower panel), in the range from 0°C to 20°C, and plotted such that each of the density axes covers the same total range of density.

Fig. 4 The seasonal progression of whole lake heat content (cal cm^{-2}; x 4.19 = Joules cm^{-2}), from January 1977 to March 1979.

Fig. 5 The seasonal progression of total stability (S_{tot}), and thermal stability (S_t: gm-cm cm^{-2}; x 9.807 10^{-5} = Joules cm^{-2}). The region between the two curves is equivalent to the saline stability (S_s).

Fig. 6 Stability due to the salinity gradient (S_s: gm-cm cm^{-2}), calculated as $S_{tot} - S_t$, and plotted on an expanded scale, for the period January 1977 to March 1979.

Fig. 7 The seasonal progression of salinity (gm l^{-1}), at the lake surface (0m), and 34m below the lake surface.

Fig. 8 Representative profiles (1 $month^{-1}$) of temperature (°C) and salinity (gm l^{-1}) for 1977, 1978, and early 1979.

Fig. 1

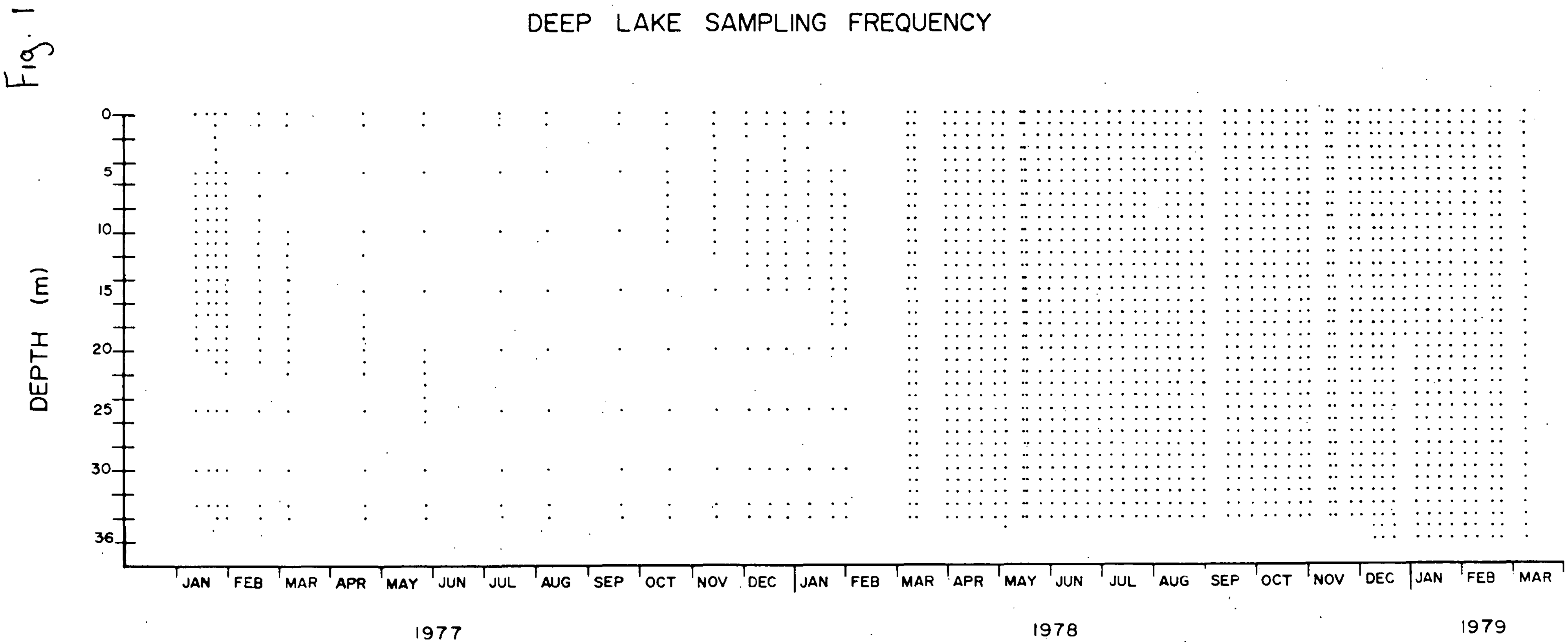
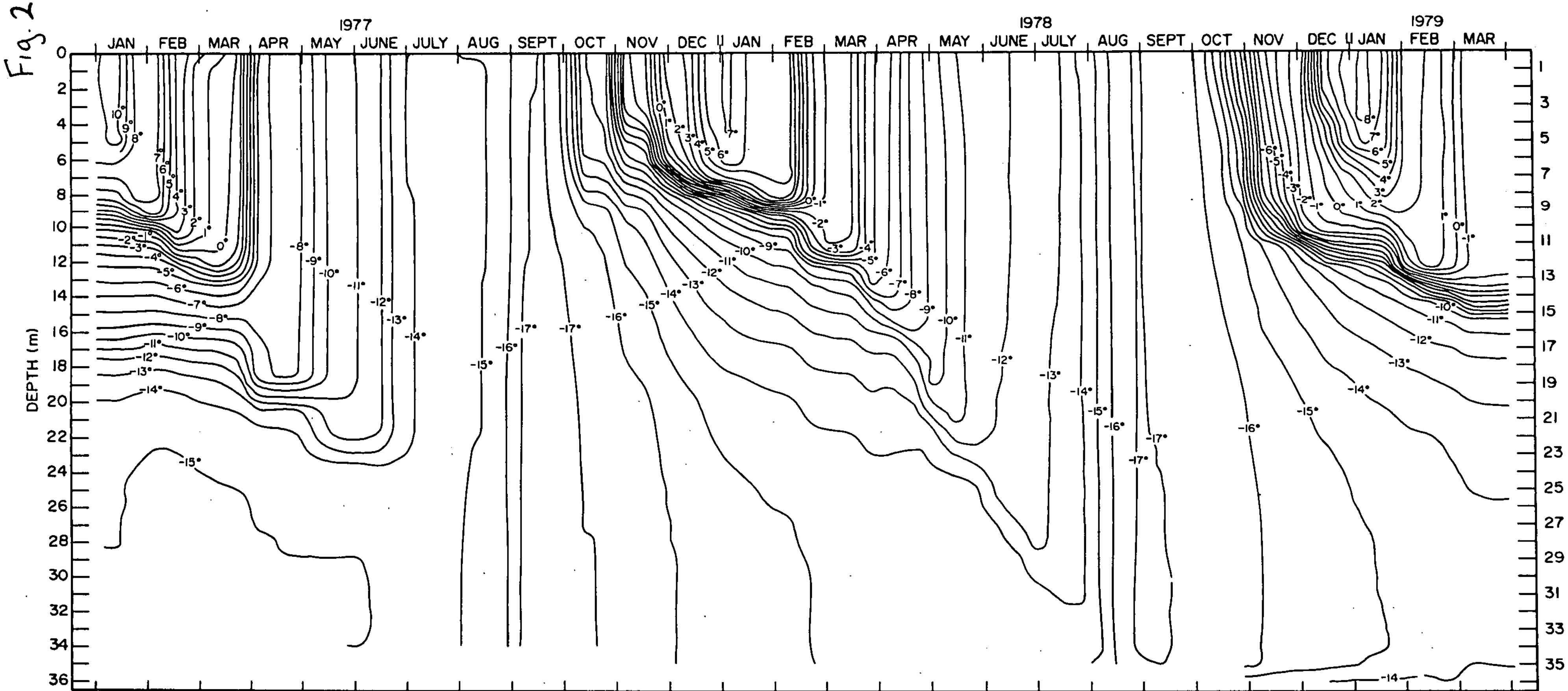
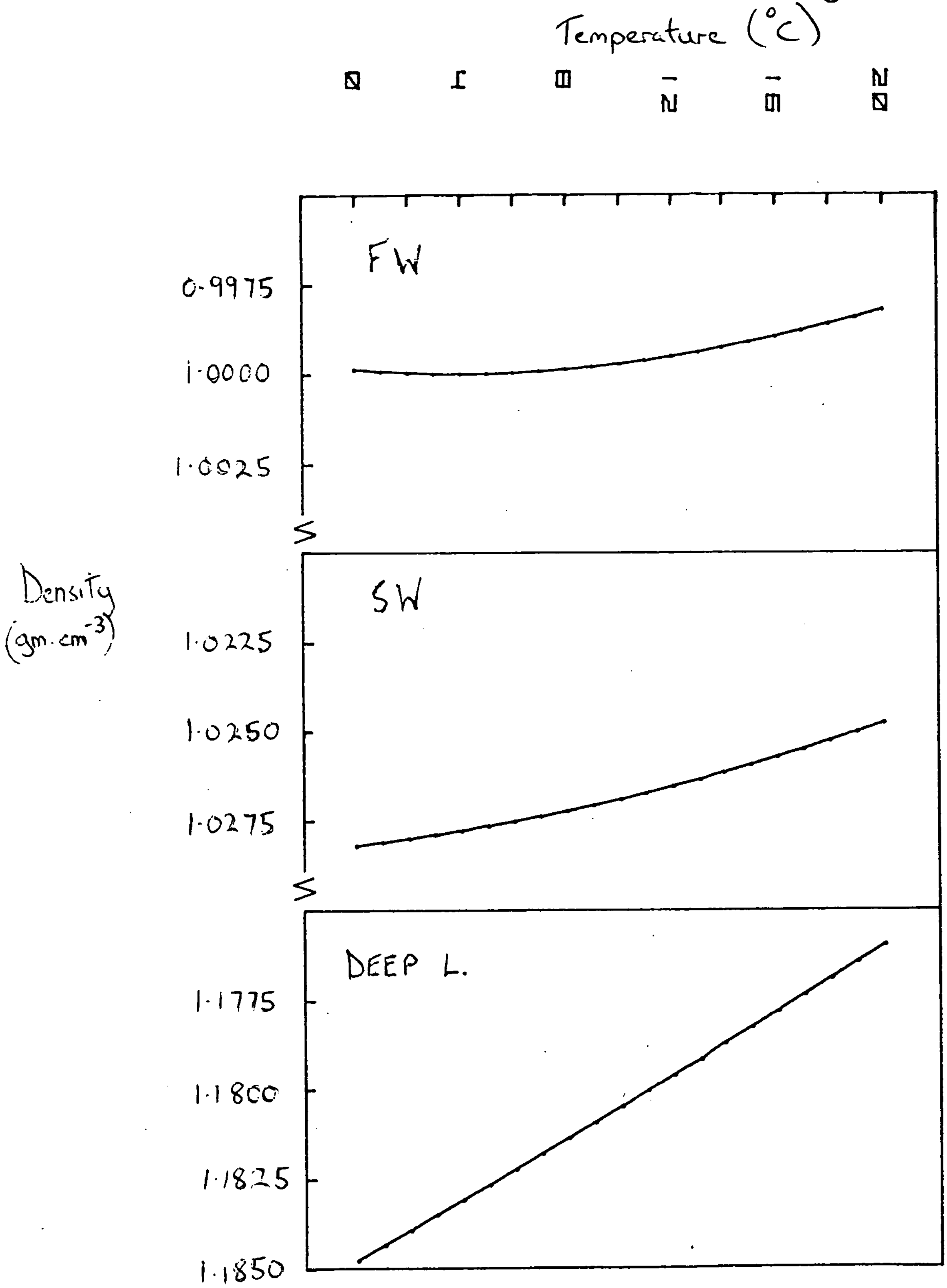

DEEP LAKE SAMPLING FREQUENCY
DEPTH (m)
0
5
10
15
20
25
30
36
JAN FEB MAR APR MAY JUN JUL AUG SEP OCT NOV DEC JAN FEB MAR APR MAY JUN JUL AUG SEP OCT NOV DEC JAN FEB MAR
1977
1978
1979

Fig. 2



1977
1978
1979
JAN FEB MAR APR MAY JUNE JULY AUG SEPT OCT NOV DEC JAN FEB MAR APR MAY JUNE JULY AUG SEPT OCT NOV DEC JAN FEB MAR
DEPTH (m)

Fig. 3.


Temperature (°C)
FW
0·9975
1·0000
1·0025
SW
1·0225
1·0250
1·0275
DEEP L.
1·1775
1·1800
1·1825
1·1850
Density
(gm.cm^{-3})

Figs 4-5

Fig 4

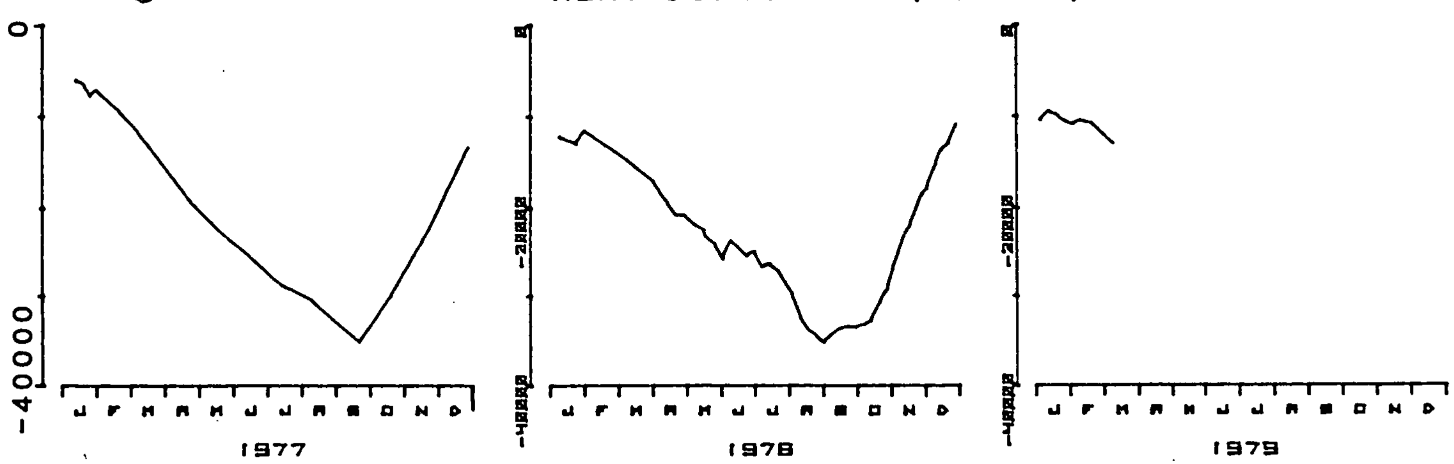
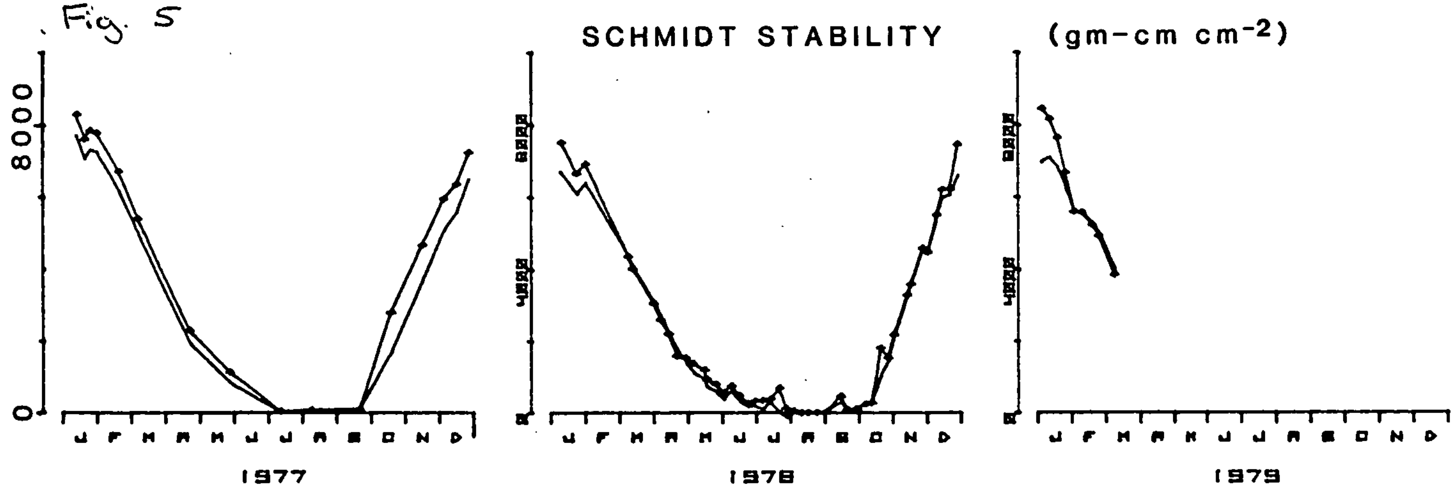


Fig. 5



Figs 6-7

Fig. 6

SALINE STABILITY gm-cm cm^{-2}

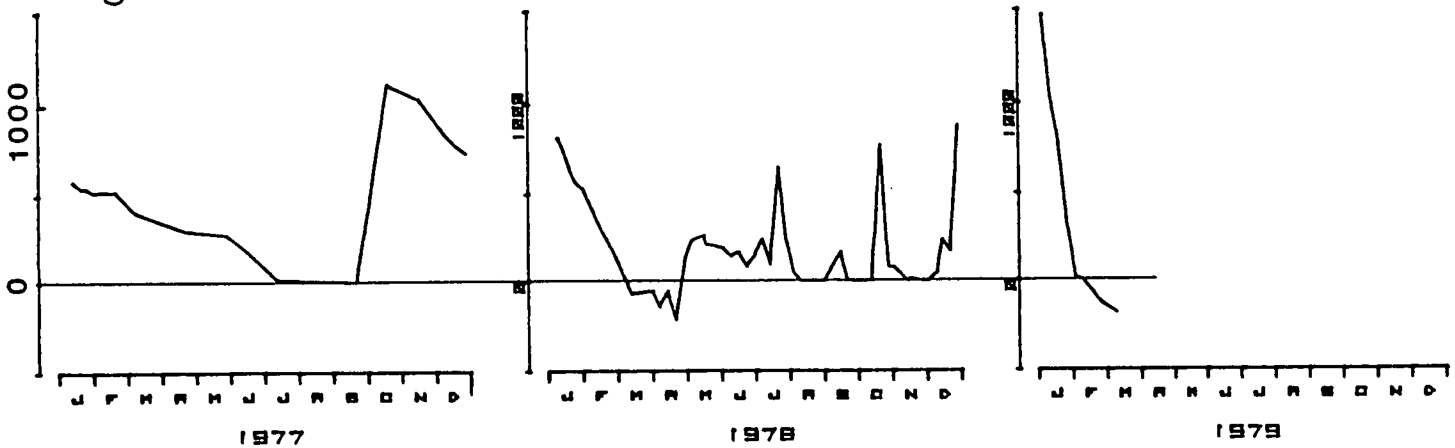


Fig. 7

SALINITY (gm/l)

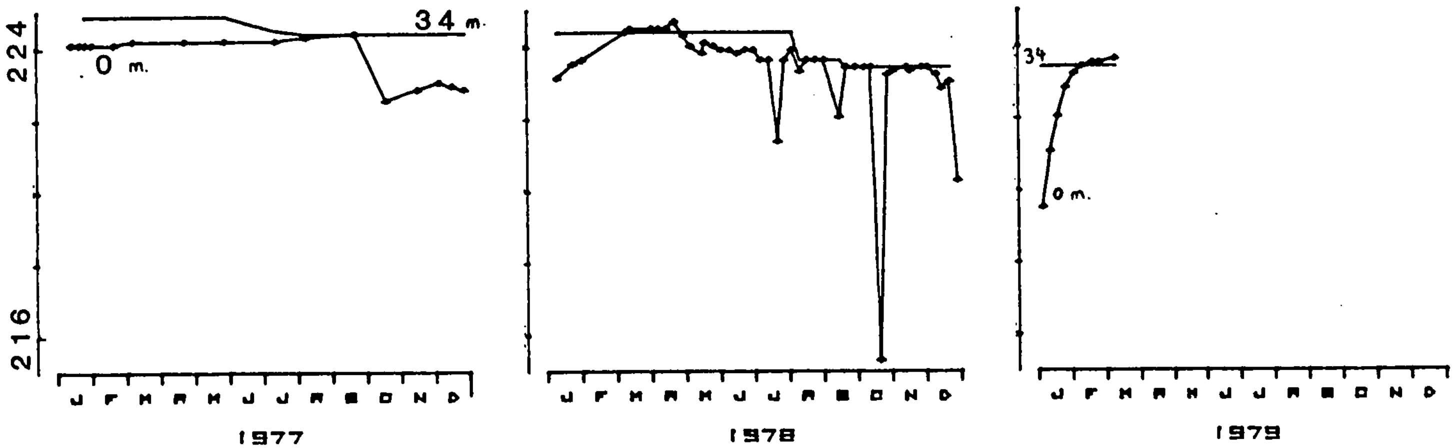


Fig. 8.

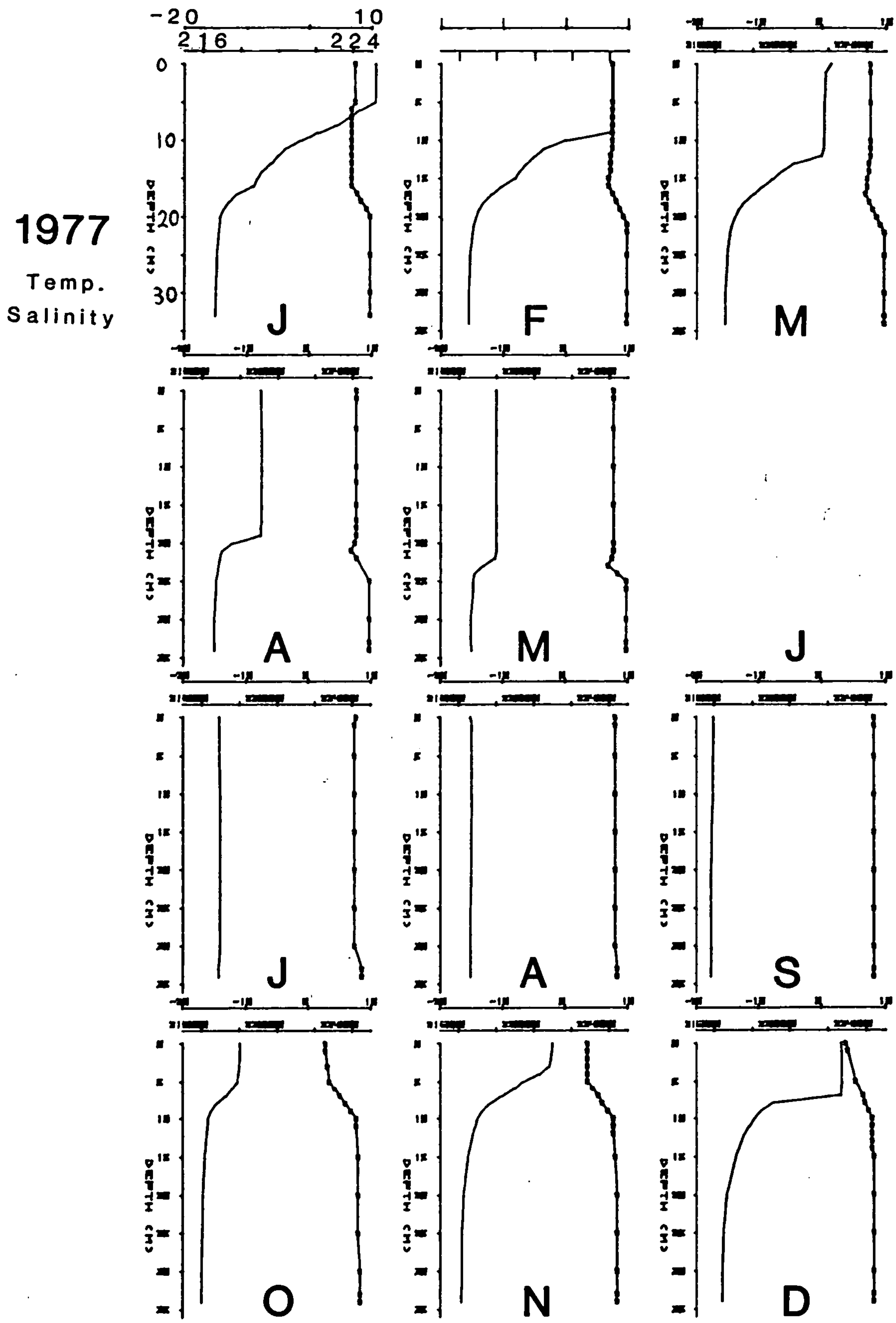

1977
Temp.
Salinity
-20
10
216
224
0
10
20
30
DEPTH (M)
J
F
M
A
M
J
J
A
S
O
N
D

Fig. 8

1978

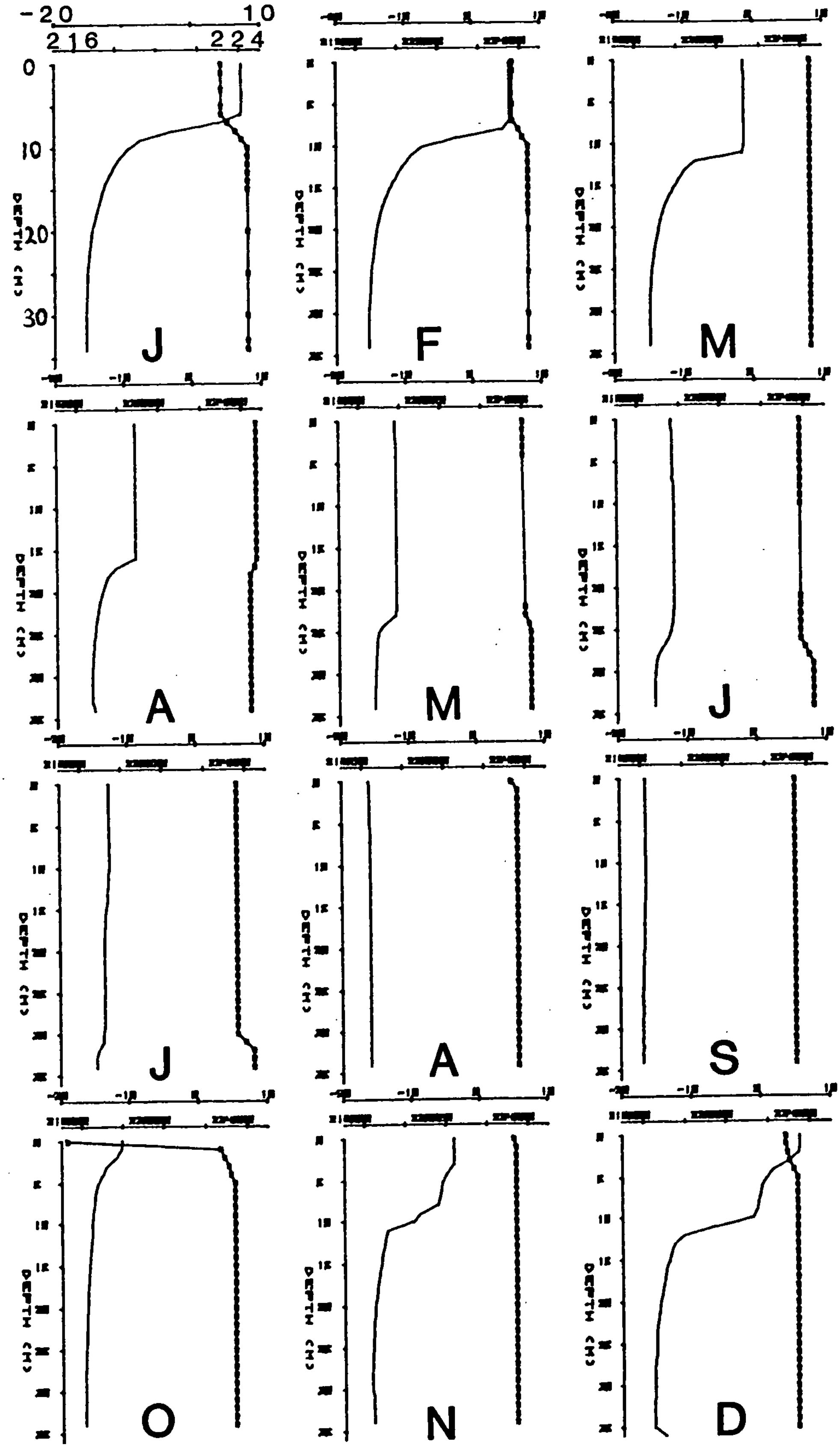
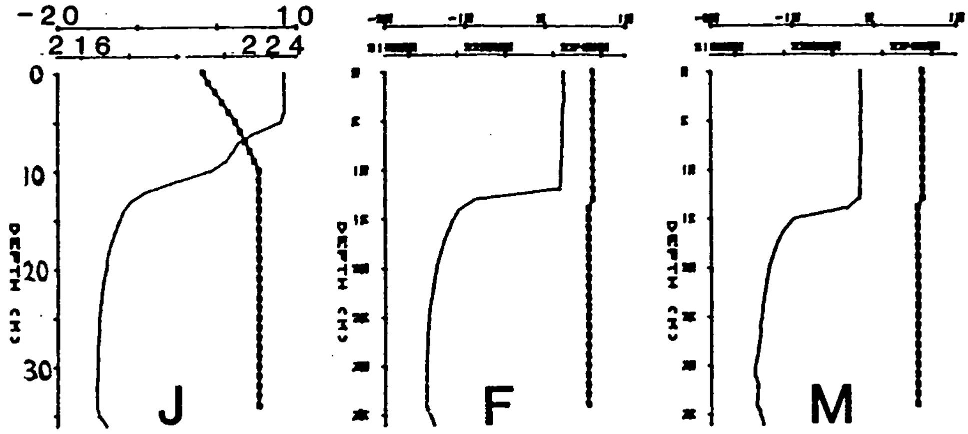


Fig. 8



1979

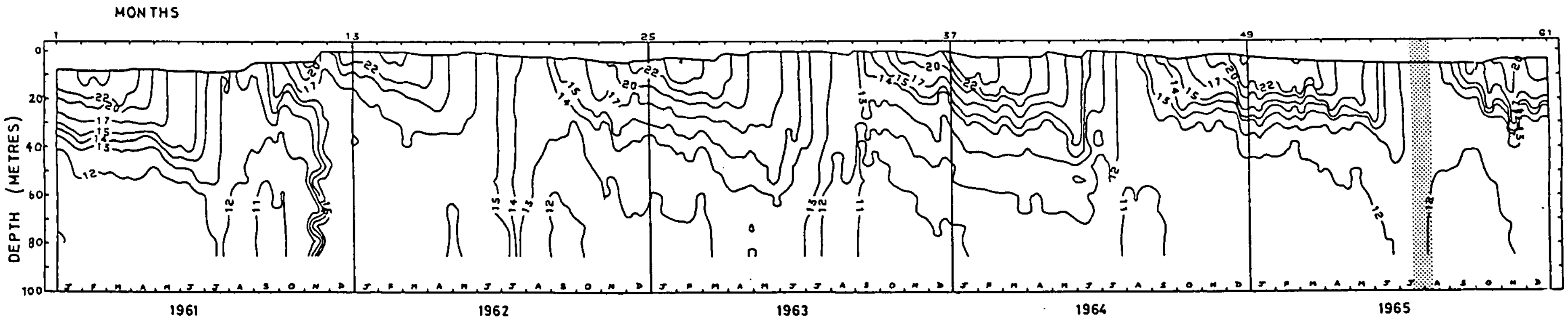
APPENDIX 4

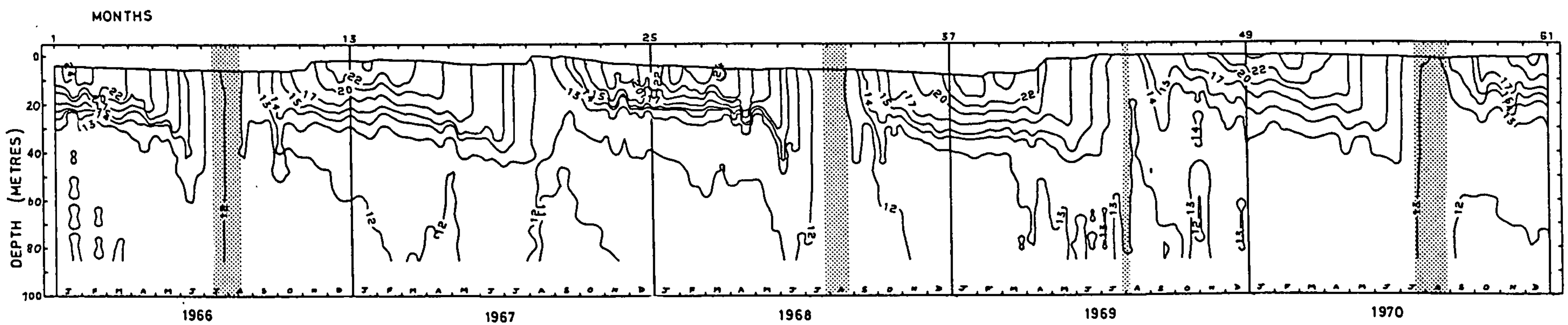
This appendix contains transparencies of some major figures in the text, which can be used as overlays for comparing information between chapters.

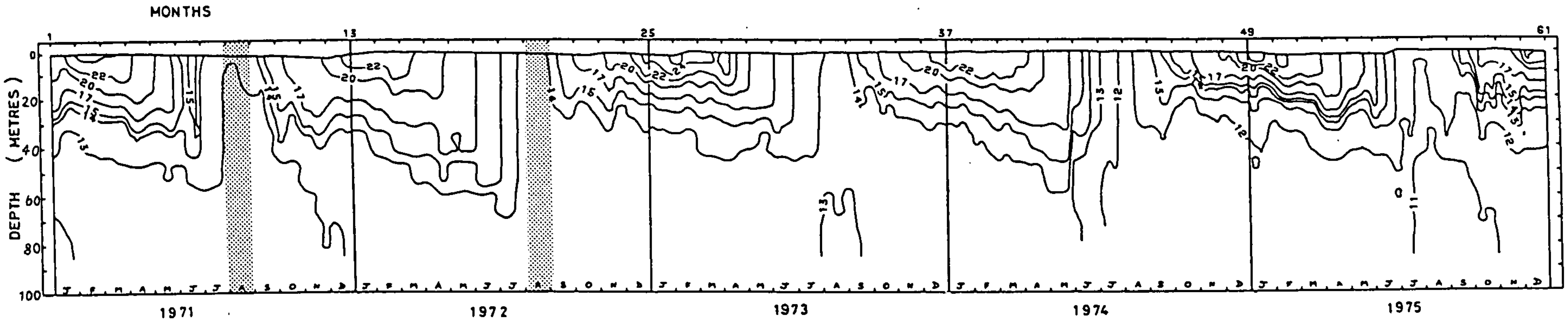
The following figures are included:-

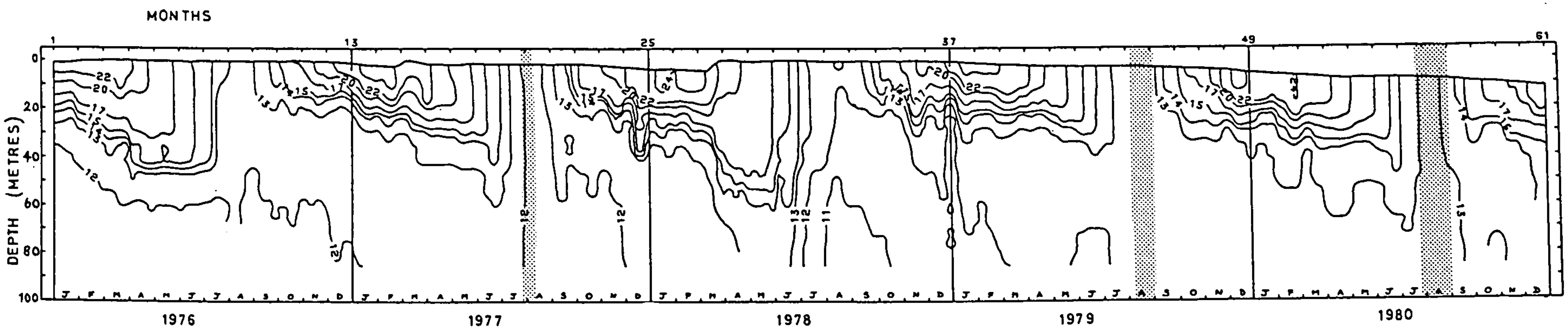
Chapter	Figure	Description
3	3.1	Temperature isopleths (°C)
3	3.5	Heat content, Stability and Birgean work (from 20 yr. temp.)
3	3.6	Monthly total inflow (minus evaporation: $\times 10^6$ m^3)
4	4.10	Monthly total outflow (HEPS: $\times 10^6$ m^3)
5	5.4	WET/DRY monthly mean temperature profiles (°C)
5	5.5	WET/DRY Mean temperature difference profiles (°C $month^{-1}$)
7	7.5	Monthly mean total phosphorus; 0 - 4.5 m sample. (mg m^{-3})

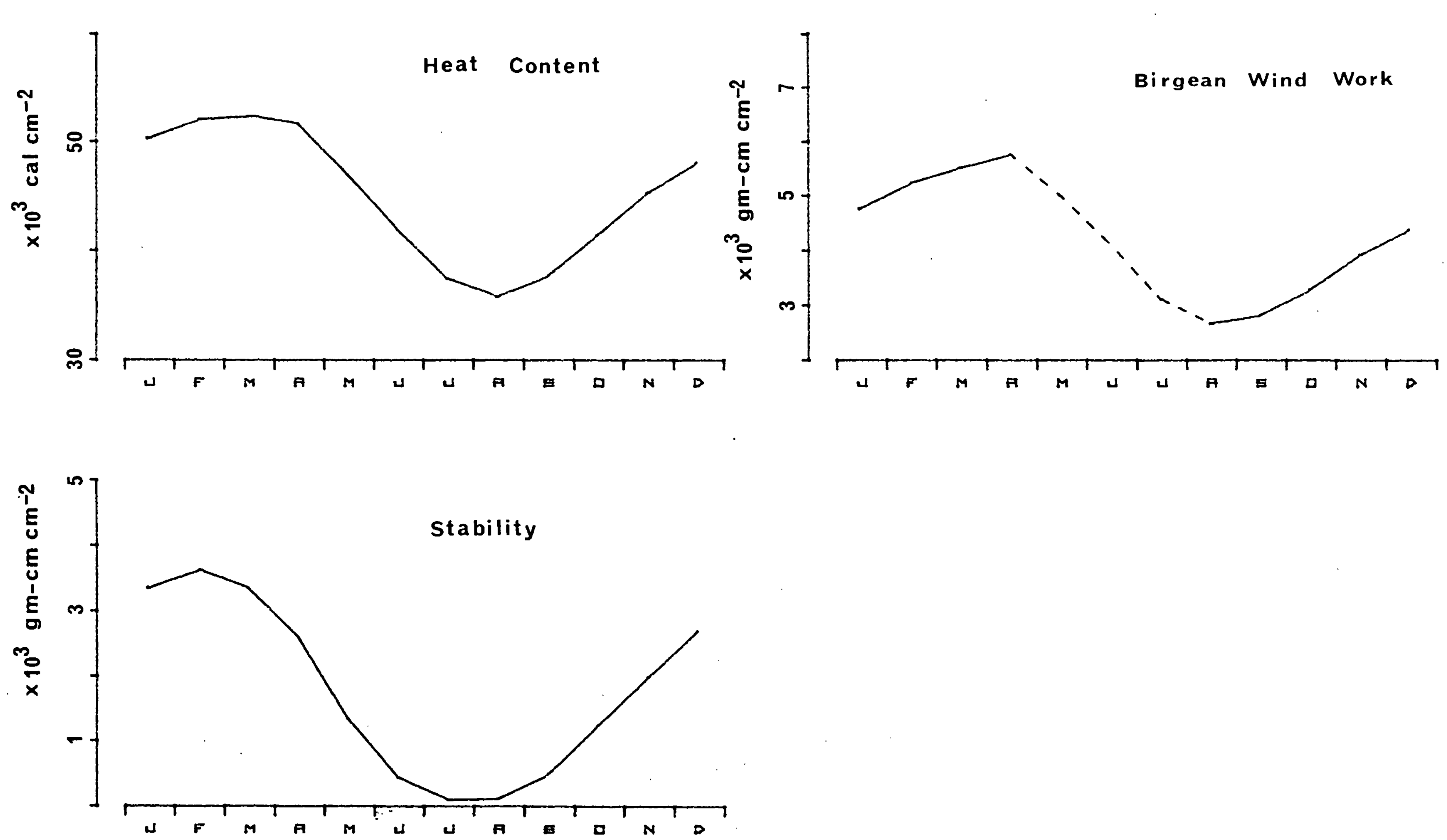
Table 3.3 (the inflow register) is also included, because it is frequently back-referenced in the later chapters.


MONTHS
1
13
25
37
49
61
DEPTH (METRES)
0
20
40
60
80
100
J F M A M J J A S O N D
1961
1962
1963
1964
1965

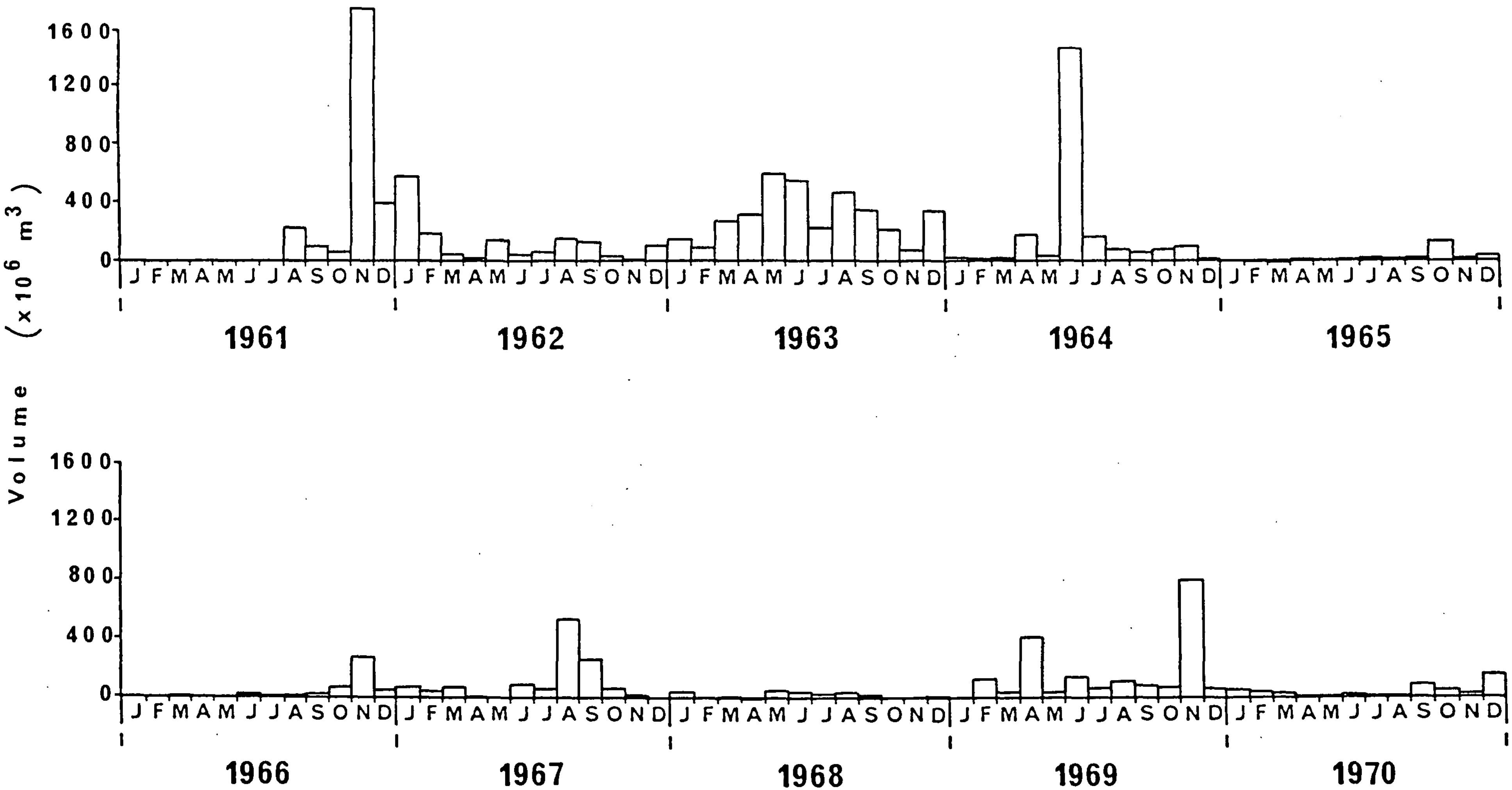

MONTHS
1
13
25
37
49
61
DEPTH (METRES)
0
20
40
60
80
100
J F M A M J J A S O N D
1966
1967
1968
1969
1970

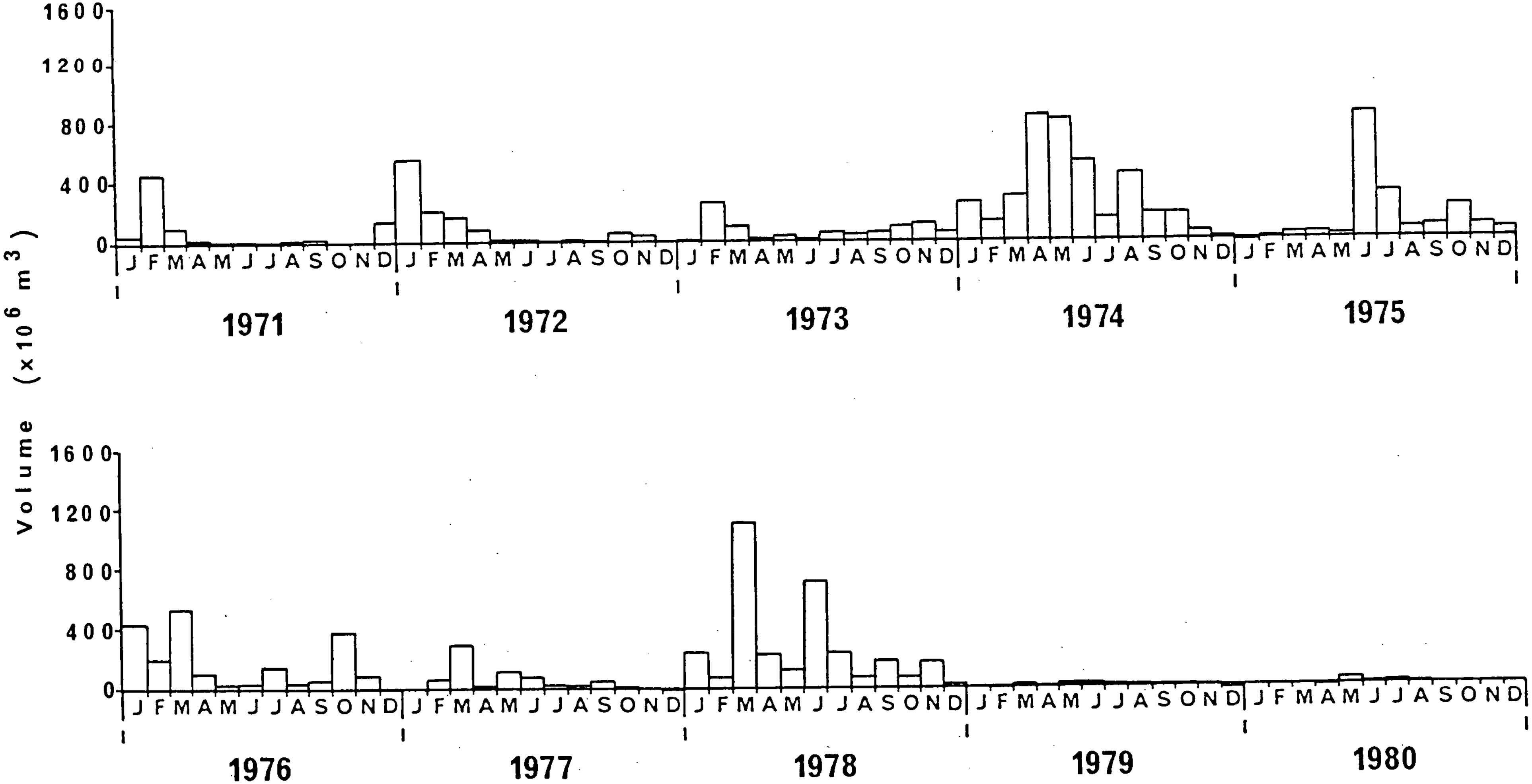

MONTHS
DEPTH (METRES)
1971
1972
1973
1974
1975


MONTHS
DEPTH (METRES)
1976
1977
1978
1979
1980

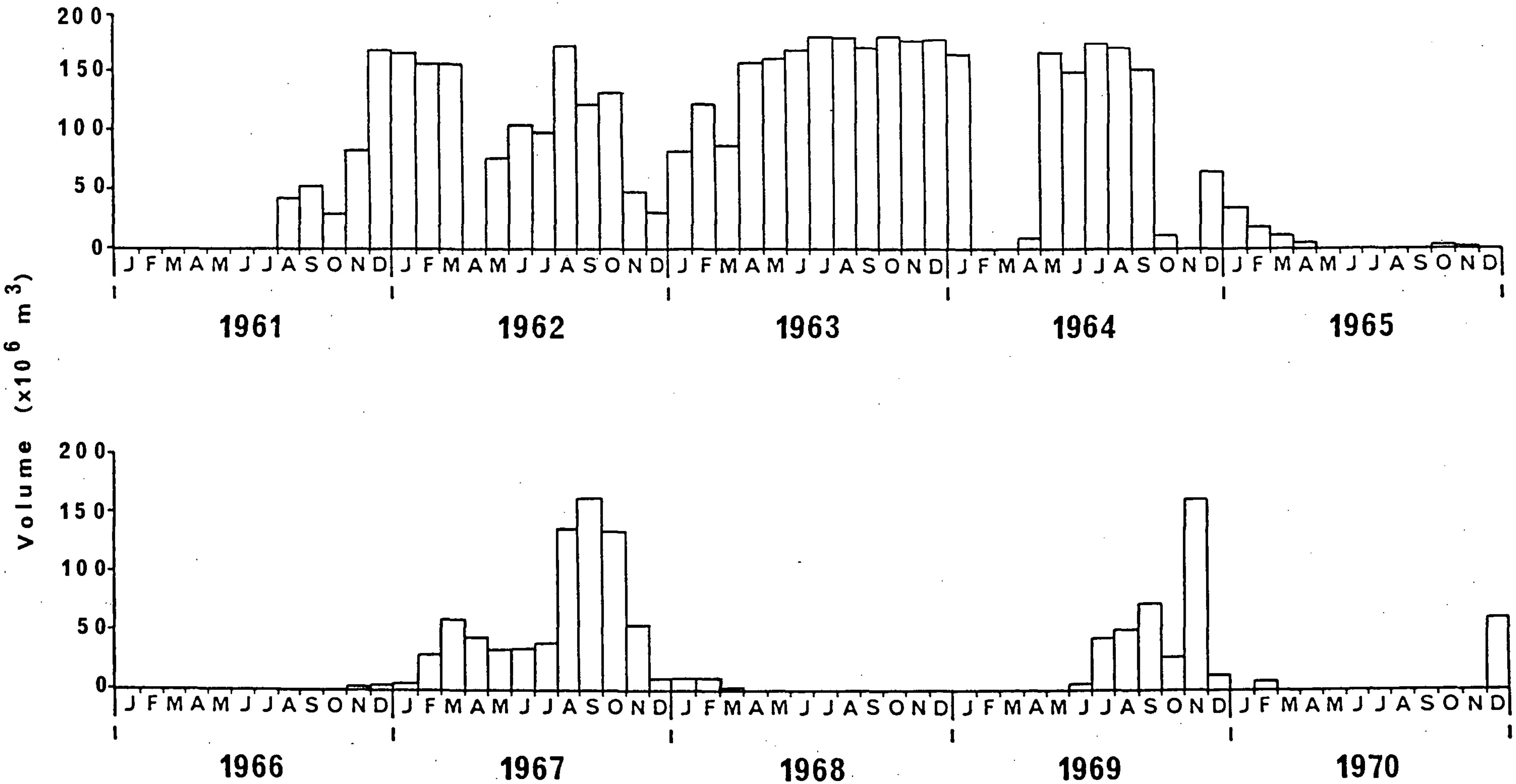


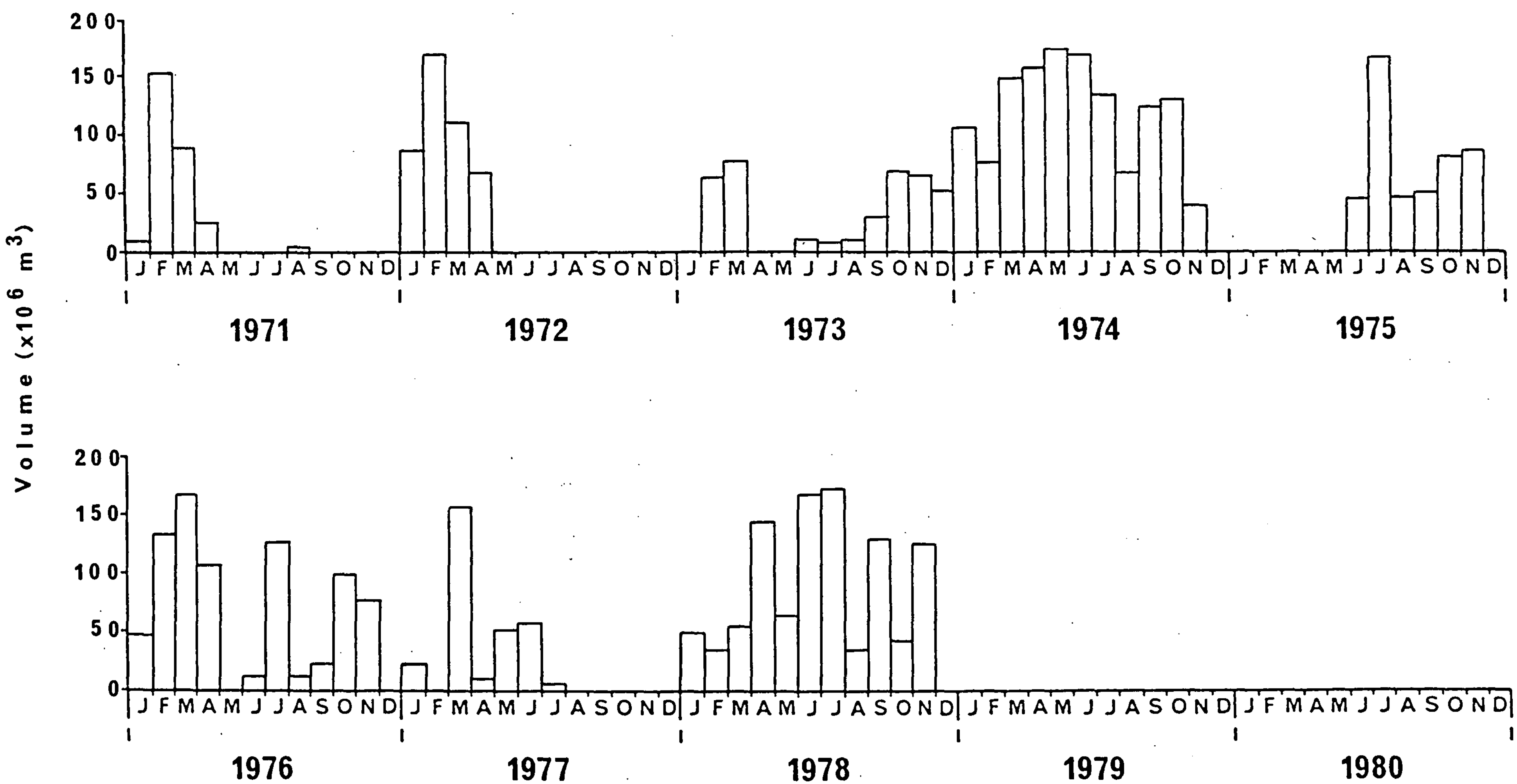
Heat Content
$\times 10^3$ cal cm^{-2}
30
50
J F M A M J J A S O N D
Birgean Wind Work
$\times 10^3$ gm-cm cm^{-2}
3
5
7
J F M A M J J A S O N D
Stability
$\times 10^3$ gm-cm cm^{-2}
1
3
5
J F M A M J J A S O N D


Volume (x10^6 m^3)
1600
1200
800
400
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

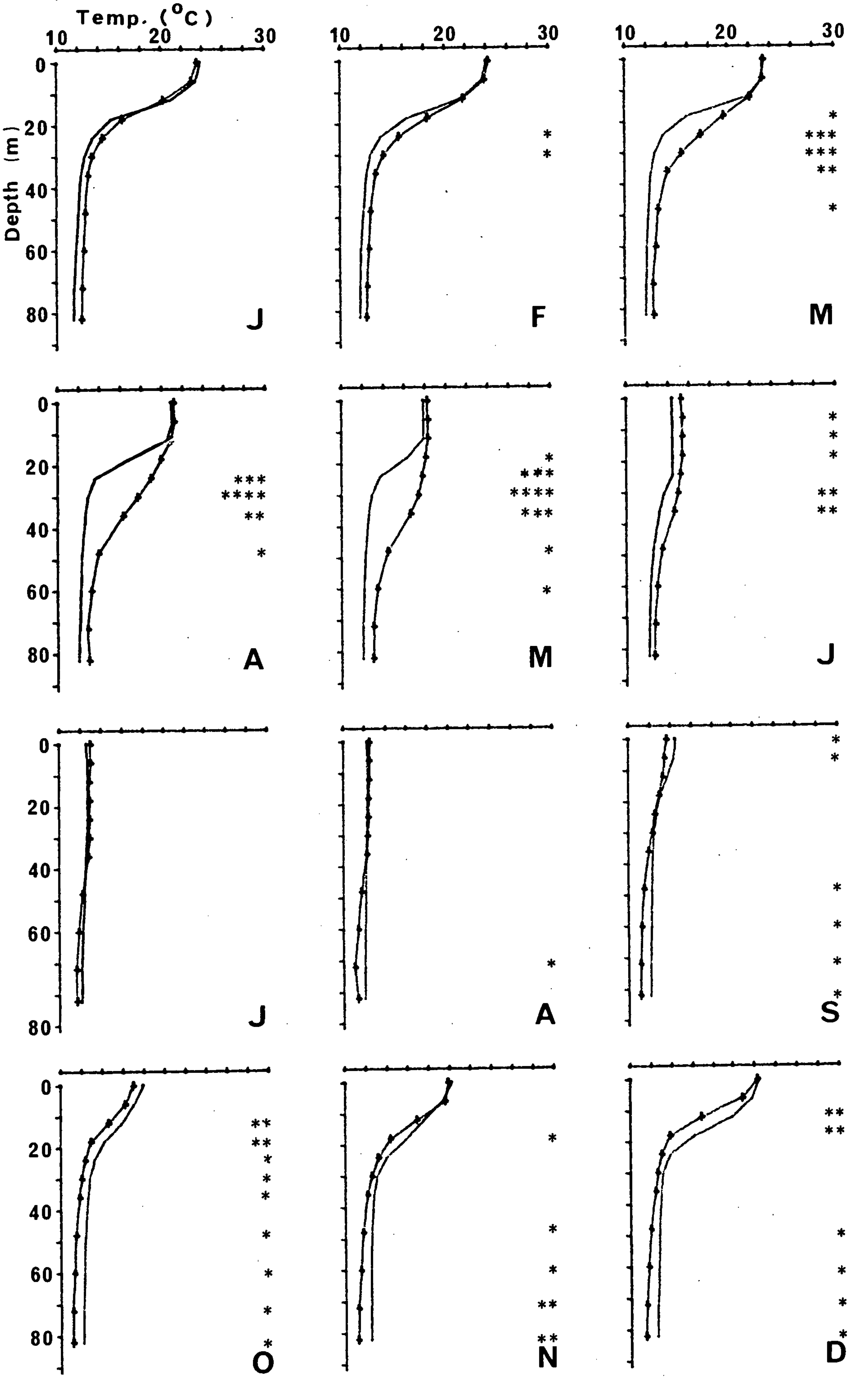


Volume (x10^6 m^3)
1600
1200
800
400
0
J F M A M J J A S O N D
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980

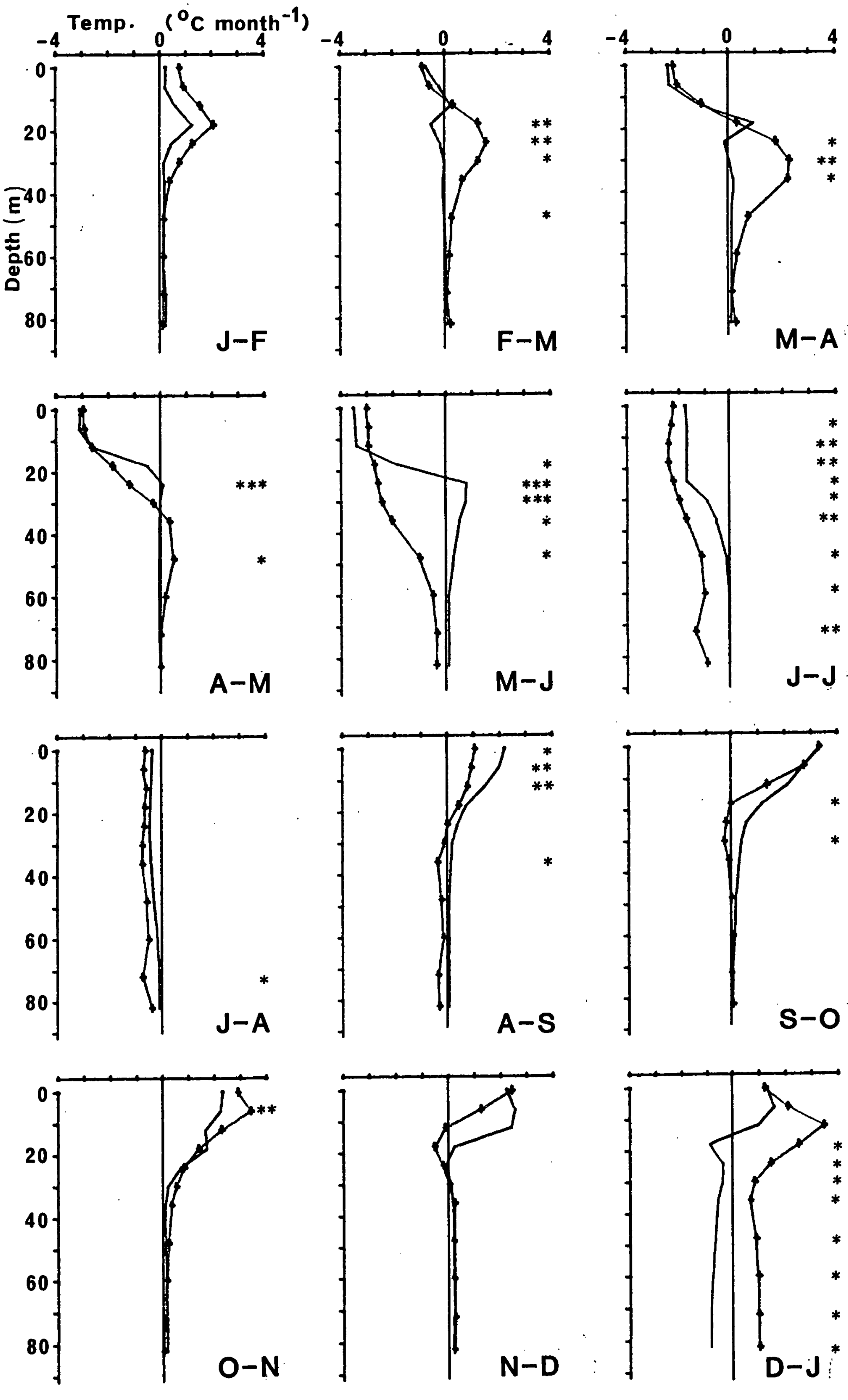

Volume (x10^6 m^3)
200
150
100
50
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970



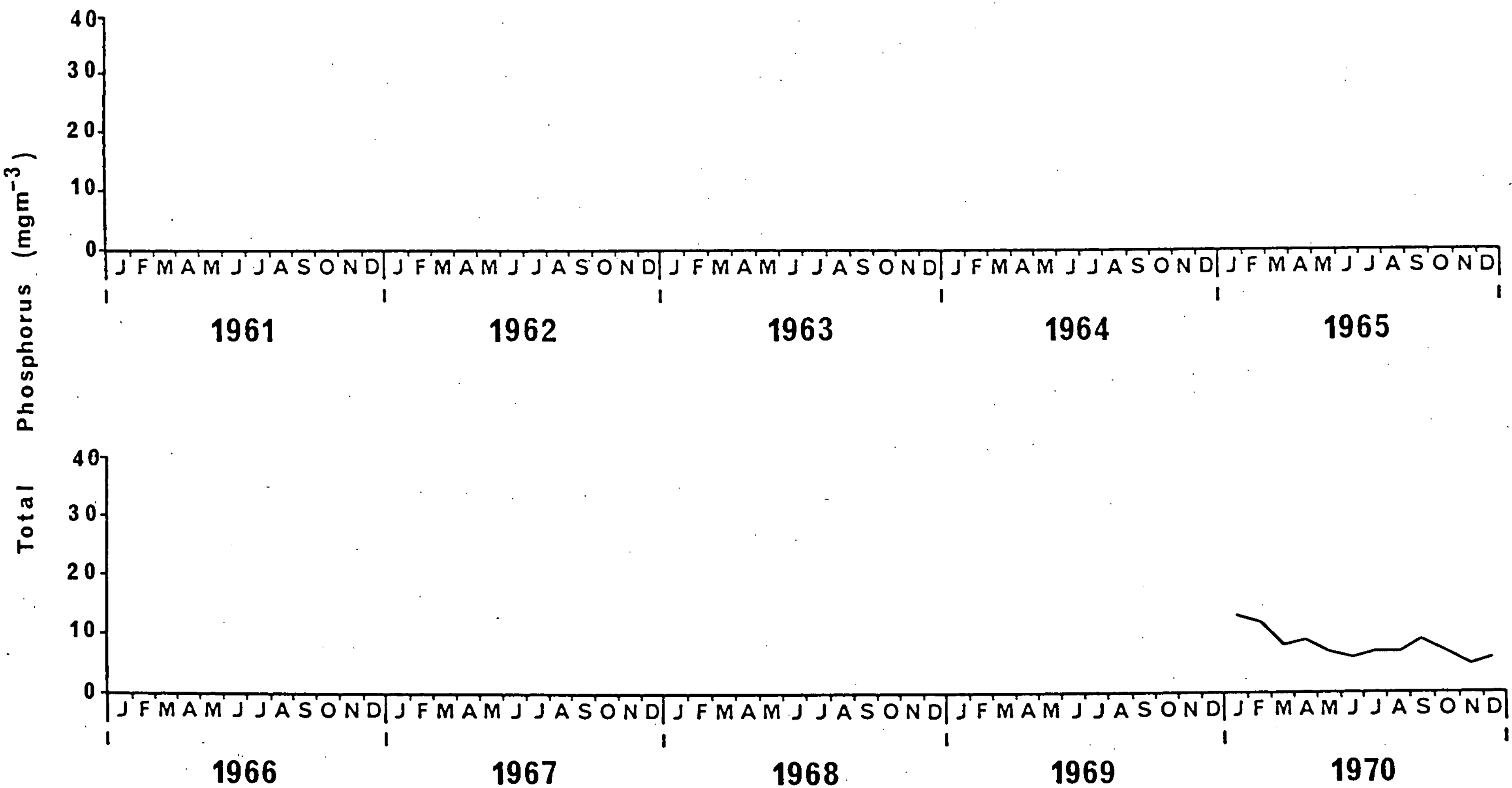
Volume (x10^6 m^3)
200
150
100
50
0
J F M A M J J A S O N D
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980



Temp. (°C)
10
20
30
Depth (m)
0
20
40
60
80
J
F
M
A
M
J
J
A
S
O
N
D



Temp. (°C month⁻¹)
-4
0
4
Depth (m)
0
20
40
60
80
J-F
F-M
M-A
A-M
M-J
J-J
J-A
A-S
S-O
O-N
N-D
D-J


Total Phosphorus (mgm^{-3})
40
30
20
10
0
J F M A M J J A S O N D
1961
1962
1963
1964
1965
1966
1967
1968
1969
1970

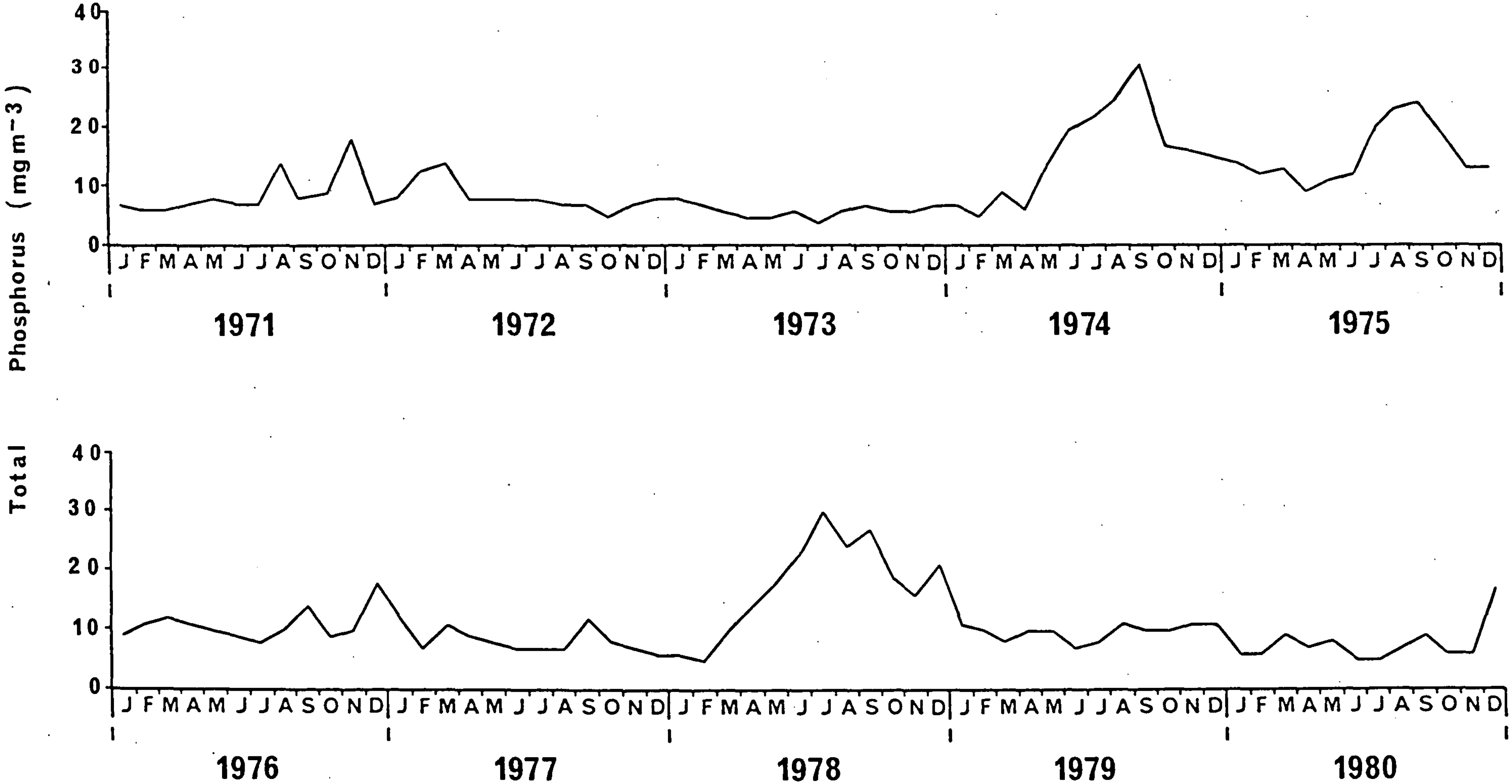

Total Phosphorus (mgm⁻³)
40
30
20
10
0
J F M A M J J A S O N D
1971
1972
1973
1974
1975
1976
1977
1978
1979
1980

Table 3.3 Inflow Register, 1961 - 1980. A list of inflows that were detectable by their individual effect on the water quality, at site 3D. The procedure used to select these inflows is described in the Methods, Chapter 2.

Inflow Initiation Date	Date of Maximum Recorded Effect Turbidity	Date of Maximum Recorded Effect Chloride	Total Inflow (minus evaporation)[a] x 10^6 m^3	Inflow type	Relative Effect Turb.	Relative Effect Cl^-	Maximum Effect Turbidity Hellige units	Maximum Effect Chlor. mg l^{-1}
3/61	4/4/61	11/4/61		U	+	+	15	25
7/61	8/8/61	22-30/8/61		U	+	+	25	34
25/8/61	5/9/61	12/9/61	198.4	U	+	-	11	24
16/11/61	24/11/61	30/11/61	1582.7	U	+	-	700	7
12/7/62	26/7/62	26/7/62	40.8	U	-	+	9	17
14/8/62	4/9/62	11/9/62	131.5	U	-	+	7	24
29/4/63	14/5/63	28/5/63	597.1	I	+	-	40	11
5/6/63	18/6/63		446.5	I	+		30	18[b]
26/6/63	9/7/63	2/7/63	94.3	U	-	-	9	11
12/7/63	23/7/63	23/7/63	123.8	U	+	+	20	32
27/8/63	10/9/63	10/9/63	564.0	U	+	-	140	16
9/12/63	17/12/63		264.4	I	+		9	18[b]
10/6/64	16/6/64	16/6/64	1346.2	U	+	-	220	6
16/7/64	11/8/64		128.0	U	-		11	14[b]
8/11/66	22/11/66	15/11/66	261.9	I	+	+	4	25
6/8/67	22/8/67	15/8/67	493.4	U	+	-	250	9
15/4/69	28/4/69		340.8	I	+		9	25[b]
13/11/69	24/11/69	1/12/69	545.0	I	+	-	8	17
1/9/70	22/9/70		57.9	U	+		10	13[b]
9/2/71	15/2/71	15/2/71	173.2	I	+	-	13	16
14/1/72	7/2/72	6/3/72	552.2	I	+	-	8	13
4/3/72	20/3/72		141.9	I	+		15	20[b]
26/2/73	5/3/73	5/3/73	112.2	I	+	+	6	24
11/7/73	23/7/73	6/8/73	35.0	U	+	-	3	13
11/3/74	18.3.74		188.9	I	+	-	8	15[b]
20/4/74	13/5/74	13/5/74	589.7	I	+	-	15	14
26/5/74	5/6/74	18/6/74	812.1	U	+	+	45	20
27/8/74	9/9/74		515.7	U	+	-	80	17
21/6/75	30/6/75	30/6/75	834.1	U	+	-	200	11
15/7/75	4/8/75	28/7/75	125.5	U	+	+	16	25
23/2/76	8/3/76	8/3/76	466.3	I	+	-	13	11
17/10/76	1/11/76	22/11/76	356.7	U	+	+	40	23
28/2/77	7/3/77	7/3/77	216.4	I	+	-	12	14
18/3/78	28/3/78	3/4/78	1083.0	I	+	-	100	14
10/4/78	17/4/78	24/4/78	119.6	I	+	-	100	14
1/6/78	19/6/78	19/6/78	562.8	I	+	+	70	12
20/6/78		17/7/78	323.5	U		+	20[b]	23

[a] From the initiation of the inflow to the date of maximum recorded turbidity.

[b] No obvious change in this parameter, so the ambient value at the depth and time of the maximum effect for the other parameter is reported.