UNITED NATIONS ENVIRONMENT PROGRAMME

REGIONAL SEAS

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Physical ocean environment in the South Pacific Commission Area

UNEP Regional Seas Reports and Studies No. 83

Prepared in co-operation with

SPC SPEC ESCAP

UNEP 1986

PREFACE

Fourteen years ago the United Nations Conference on the Human Environment (Stockholm 5-16 June 1972) adopted the Action Plan for the Human Environment, including the General Principles for Assessment and Control of Marine Pollution. In the light of the results of the Stockholm Conference, the United Nations General Assembly decided to establish the United Nations Environment Programme (UNEP) to "serve as a focal point for environmental action and co-ordination within the United Nations system" (General Assembly resolution 2997(XXVII) of 15 December 1972). The organizations of the United Nations system were invited "to adopt the measures that may be required to undertake concerted and co-ordinated programmes with regard to international environmental problems", and the "intergovernmental and non-governmental organizations that have an interest in the field of the environment" were also invited "to lend their full support and collaboration to the United Nations with a view to achieving the largest possible degree of co-operation and co-ordination". Subsequently, the Governing Council of UNEP chose "Oceans" as one of the priority areas in which it would focus efforts to fulfill its catalytic and co-ordinating role.

The Regional Seas Programme was initiated by UNEP in 1974. Since then the Governing Council of UNEP has repeatedly endorsed a regional approach to the control of marine pollution and the management of marine and coastal resources and has requested the development of regional action plans.

The Regional Seas Programme at present includes ten regions[1/] and has over 120 coastal States participating in it. It is conceived as an action-oriented programme having concern not only for the consequences but also for the causes of environmental degradation and encompassing a comprehensive approach to combating environmental problems through the management of marine and coastal areas. Each regional action plan is formulated according to the needs of the region as perceived by the Governments concerned. It is designed to link assessment of the quality of the marine environment and the causes of its deterioration with activities for the management and development of the marine and coastal environment. The action plans promote the parallel development of regional legal agreements and of action-oriented programme activities[2/].

The idea for a regional South Pacific Environment Management Programme came from the South Pacific Commission (SPC) in 1974. Consultations between SPC and UNEP led, in 1975, to the suggestion of organizing a South Pacific Conference on the Human Environment. The South Pacific Bureau for Economic Co-operation (SPEC) and the Economic and Social Commission for Asia and the Pacific (ESCAP) soon joined SPC's initiative and UNEP supported the development of what became known as the South Pacific Regional Environment Programme (SPREP) as part of its Regional Seas Programme.

1/ Mediterranean, Kuwait Action Plan Region, West and Central Africa, Wider Caribbean, East Asian Seas, South-East Pacific, South Pacific, Red Sea and Gulf of Aden, Eastern Africa and South Asian Seas.

2/ UNEP: Achievements and planned development of UNEP's Regional Seas Programme and comparable programmes sponsored by other bodies. UNEP Regional Seas Reports and Studies No. 1. UNEP, 1982.

An Action Plan for the South Pacific Regional Environment Programme (SPREP) was adopted at the Conference on Human Environment in the South Pacific at Rarotonga, 8-11 March 1982, and was endorsed seven months later at the South Pacific Conference and South Pacific Forum[3/].

This document has been prepared by Mr. Bruno Wauthy of ORSTOM (Office de la Recherche Scientifique et Technique Outre-Mer) as a contribution to the South Pacific Regional Environment Programme. The sponsors of the study would like to express their gratitude to the author and ORSTOM.

3/ SPC/SPEC/ESCAP/UNEP: Action Plan for managing the natural resources and environment in the South Pacific Region. UNEP Regional Seas Reports and Studies No. 29. UNEP, 1983.

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GEOGRAPHICAL INTRODUCTION TO THE SPC AREA

Geographical extent

The South Pacific Commission (SPC) Area (Fig. 1B) extends from 130°W to 130°E, on either side of the international dateline, between the two tropics. It is unequally distributed as regards latitude, with two thirds of its area in the Southern Hemisphere and one third in the Northern Hemisphere. Its eastern area corresponds roughly to the central South Pacific, and its western portion to the western Pacific, excluding the areas of the Coral Sea to the South and the Philippine Sea to the North which wash the Australian and Philippines coastlines respectively.

The area covered is approximately 27.5 million km^2 and represents four-tenths of the intertropical Pacific (Fig. 1A) and almost one-fifth of the area of all oceans within the intertropical belt.

The distances involved are considerable, measuring 8,000 km from Guam to Tahiti; 11,000 km from Pitcairn to Palau; 5,500 km from the North of the Marianas to Noumea and 3,600 km from Christmas Island to Rapa. From Papeete (Tahiti), the nearest continental landfalls are Sydney at a distance of 6,000 km; San Francisco at 6,600 km; Terre Adélie at 7,000 km; Vladivostok at 7,500 km; Valparaiso at 7,700 km and Panama at 8,200 km.

The SPC Area is thus primarily a vast ocean expanse covering the major portion of the area known to geographers as Oceania.

Emerged land

The emerged land takes the form of a multitude (poly) of small (micro) islands (nesos) scattered over a vast expanse of ocean and mostly grouped in archipelagos which form Micronesia in the North-west and Polynesia in the East. They contrast with the larger islands of Melanesia, which extend from Fiji to New Guinea and form part of the structural mass of the Australian continent.

Table I shows:

The areas of islands or archipelagos (taken from Le Million, Vol. XV, OCEANIE, ed. Grange Batelière, Paris) in Micronesia, Polynesia and Melanesia, excluding, in the last case, Papua-New Guinea, in view of its clearly continental character;

The areas of ocean occupied by these major provinces, estimated on the basis of the geographical distribution of the archipelagos;

Ratings on an "insularity index" expressed as percentages and calculated on the basis of the ratio of emerged land areas to ocean area.

The insularity characteristic is extreme in Micronesia where the islands represent only an infinitesimal fraction of the ocean area; it is very pronounced in Polynesia where the remoteness from continents accentuates the impression of isolation, but is less marked in island Melanesia, where there is almost 100 times as much land area as in Micronesia.

Nevertheless, throughout the SPC Area, insularity remains the predominant characteristic, since only 2 per cent of the area consists of emerged land, whereas water represents 98 per cent. Figure 1A, which could be regarded as an "artist's view of Earth from space directly over Jarvis Island", illustrates this clearly.

While the presence of the small islands has only a marginal influence on the characteristics of the ocean environment, the large islands of Melanesia, some of which are very high (chains with peaks of over 4,000 metres in New Guinea) are important in that their heavy rainfall produces considerable erosion, leaching large quantities of soluble minerals (nutrient salts and trace elements) throughout the year, thereby increasing the fertility of the waters of the adjacent ocean. The highest rates of chemical leaching from the rocks are in fact seen in the humid equatorial zone where temperatures are high and rainfall abundant (Strakhov, 1960).

The ocean floor (Fig. 2)

The relief of the ocean floor is very uneven in the western part of the area where the continental margins are found.

The continental shelf down to the 200 m isobath demarcates the continent proper and illustrates clearly that New Guinea is directly linked to Australia. Most of the Arafura Sea is less than 100 metres in depth, and the Torres Strait less than 20 metres.

The margin area extends from the continent to the island arcs in the East which form an almost continuous chain of islands or shallows stretching from New Zealand, through Tonga, the Solomon Islands, the Bismarck Archipelago, Palau and the Marianas to Japan. This is the area of such adjacent seas as the Tasman Sea, the Coral Sea, the Solomon Sea and the Philippine Sea. Inside the margin area seamounts support archipelagos such as Fiji, Vanuatu, New Caledonia and the Loyalty Islands, while ridges more than 2,000 metres deep (Lord How Rise, Norfolk Island Ridge and Lau Ridge) isolate plateaux where the depth exceeds 2,000 metres (Northern Fiji and Chesterfield) and even 4,000 metres (New Guinea Basin, Solomons Basin, New Hebrides Basin, Fiji Basin and Coral Sea Basin).

The island arcs are bounded on the East by deep trenches, the most distinctive of which are the Marianas Trench (10,915 metres), the Palau Trench (8,050 metres), the Bougainville Trench (9,103 metres), the New Hebrides Trench (9,165 metres), the Tonga Trench (10,882 metres) and the Kermadec Trench (10,047 metres); in many cases the walls of these trenches are in the form of steps.

To the East of the island arcs, the relief is less accentuated. Average depth is greater than 4,000 metres; seamounts in the form of ocean ridges stretch over considerable distances. Their width varies from 200 km to 500 km; they are more or less parallel, running in a NW-SE direction, and they sometimes form basements of atolls (Carolines, Marshalls, Kiribaki and Tuvalu, Tokelau-Cook, Line Islands and Tuamotu-Gambier). In conjunction with a number of East-west transverse plateaux, they demarcate basins of various sizes (Carolines Basin, Marshalls Basin, South-west Pacific Basin, Central Basin, North-west Basin) and a number of deep trenches. Other less extensive relief features created by volcanic activity form the basements of more isolated island groups (Society Islands, Samoa, Marquesas, Hawaii).

The bathymetry illustrated in Figure 2 shows the Coral and Solomon Seas to be relatively isolated:

To the West by the continental land mass of Australia and New Guinea which form a virtually continuous barrier; to the East and North by the island arcs of the New Hebrides, the Santa Cruz, the Solomons and New Britain; to the South by shoals between New Caledonia, the Chesterfield Plateau and the Great Barrier Reef which render virtually impossible any exchange with the Tasman Sea at depths greater than 3,000 metres.

Table 1: AREAS AND DEGREE OF "INSULARITY" IN THE SPC AREA

LAND (km^2)		OCEAN ($10^6 km^2$)	INSULARITY
Marianas	404		
Marshall	181		
Palau	478		
Ponape	334		
Carolines			
Yap	216		
Truk	104		
Guam	549		
Nauru	21		
Kiribati	931		
MICRONESIA	3 218	8,8	0,036%
Tuvalu	26		
Wallis and Futuna	255		
Western Samoa	2 842		
American Samoa	197		
Tonga	699		
Tokelau	10		
Cook	455		
Niue	259		
Society	1 647		
Marquesas	1 274		
Tubuais	164		
Tuamotu-Gambier	915		
POLYNESIA	8 733	13,2	0,066%
New Britain	36 519		
New Ireland	8 651		
Manus-Admiralty	2 072		
Bougainville-Buka	10 500		
Solomons	28 446		
Fiji	18 272		
Vanuatu	14 763		
New Caledonia	19 058		
MELANESIA	138 281	5,5	2,51 %
Papua New Guinea	403 950		
Total MELANESIA	542 231	5,5	9,85 %
TOTAL SPC AREA	554 182	27,5	2,01 %

METEOROLOGY

Atmospheric circulation

Atmospheric pressure at ground level and low-level wind regime

Pressure distribution at sea level (Figure 3 - Queeney, 1974) is determined by the different reactions of the continents and oceans to the seasonal thermal factor of solar radiation.

In the Southern Hemisphere, the dominant influence is that of latitude. Isobars are approximately zonal; the sub-tropical high pressure belts are practically continuous, although there is some increase in pressure over Australia during the southern winter, while in summer, the North of the continent is invaded by the Indonesian equatorial low pressure zones.

In the Northern Hemisphere, the dominant influence in the extreme seasons is the continental effect. In the northern summer, the Indo-Asiatic continental thermal depression deepens, while the whole of the northern Pacific is covered by an anticyclone. In winter, on the other hand, a vast and powerful anticyclone forms over Asia, centred over the Himalaya massif, while the vast Aleutian depression zone pushes back the sub-tropical ocean anticyclone to the eastern part of the northern Pacific.

This pressure field determines the low level wind patterns. In the Central Pacific, the constancy of the sub-tropical high pressure belts (Hawaii anticyclone, Easter Island anticyclone) sustains a year-round regime of trade winds which blow steadily in the direction of the near equatorial low-pressure zone.

In the western Pacific, the influence of Asia results in a seasonal pattern of monsoons which blow from the high-pressure zones in the winter hemisphere towards the continental low-pressure zones in the summer hemisphere. In the Philippines, New Guinea and northern Australia, for example, there is an alternation of winds in the North to North-west sector during the northern winter, and of winds in the South to South-east sector during the southern winter.

Throughout a "normal year", the equatorial zone including Indonesia is covered by troughs of relatively low pressure, with the result that there is usually a decrease in the pressure gradient along the equator between the eastern and western Pacific, resulting in a zonal circulation known as the "Walker cell", consisting of a lower branch in the form of the trade winds, a rising branch over Indonesia and an upper return branch feeding subsident air into a descending branch over the eastern Pacific.

Low level wind field

An understanding of the wind field at the ocean surface is of paramount importance for sailors, oceanologists and climatologists.

Since the agitation of the ocean surface is due essentially to the wind stress on the surface, the average wind field (speed, direction) is used by oceanographers to calculate a number of derived values (wind stress, curl of the wind) needed for ocean circulation models.

Ocean-atmosphere exchanges, which determine weather to a large extent, also depend on wind speed..

Wind field can be described on the basis of observations made at weather stations established on the islands or of observations made by vessels at sea.

Winds observed on land, particularly from high islands, may be only marginally representative of the wind on the open sea (Wyrtki and Meyers, 1975); the speeds of geostrophic winds calculated on the basis of differences in atmospheric pressure from one island to another in the 20°N to 20°S belt do not give an accurate estimate of the force of the trade winds. Consequently, the wind field is usually determined on the basis of observations made from vessels (Hidaka, 1985; Hellerman, 1967; Wyrtki and Meyers, 1975; Kutsuwada and Sakurai, 1982).

The wind field in the Pacific inter-tropical zone is characterized by the trade winds. The stability of these winds makes it possible to calculate an average value for the wind stress (in $N.m^{-2}$) of the wind at the ocean surface, using the formula $\tau = \rho a\ CD\ W^2$, where ρa is the air density (more or less constant at about 1.2 $kg\text{-}m^{-3}$), CD is a friction coefficient and W the wind speed in $m.s^{-1}$.

After processing approximately 5 million wind observations collected at the Environmental Data Centre, Ashville, N.C. in Marine Deck (Tape Data Family - 11), Wyrtki and Meyers published monthly maps for the inter-tropical Pacific showing a mean wind vector (speed and direction) and a mean wind stress vector (Fig. 4), and bimonthly charts of derived values such as the curl of the wind stress or the wind divergence (grid 2° latitude x 10° longitude) (Fig. 5).

The consistency of the trade winds system is clearly illustrated in Figure 4.

North-easterly trade winds

The North-easterly trade winds are strong from November to May; they blow all the way across the Pacific to Asia where they strengthen the North-easterly winter monsoon over the China Sea. They reach as far as 5°N in the western Pacific.

The North-easterly trade winds are light from June to October, particularly in the western Pacific, where the wind stress diminishes by half. The South-westerly monsoon, which blows close to Asia in summer affects the direction of the winds as far as 170°E.

South-easterly trade winds

The South-easterly trade winds are strong from June to October, reaching their peak in July. They are lighter from December to May, particularly in the western Pacific which is then experiencing the doldrums.

To the South of a line from the Solomon Islands to the Society Islands, the direction of the trade winds is determined by high pressure zones centred near the Kermadec Islands and caused by mobile anticyclones which do not appear on mean pressure charts but represent a constant factor in day to day atmospheric dynamics in this region.

Wind-field divergence

The divergence or convergence of horizontal winds at the surface are the cause of vertical movements in the lower troposphere. Divergence is calculated on the basis of the average wind field by the formula:

$$\operatorname{div} \overline{W} = \frac{\delta Wx}{\delta x} + \frac{\delta Wy}{\delta y}$$

where Wx and Wy are the components of the average wind-speed vector $\overline{W}$ in a system of rectangular co-ordinates, and counted positively eastwards and northwards respectively.

The average annual distribution of this property is illustrated in Figure 5A, where the isolines represent intervals of $4 \times 10^{-6} S^{-1}$. The noticeable features are:

A transpacific zonal belt between 5°N and 10°N in which convergence is at its maximum. This is the Inter-tropical Convergence Zone (ITCZ) between the Northern Hemisphere trade winds and the Southern Hemisphere trade winds which have crossed the equator;

A slanting belt stretching from the equator in the western Pacific to the southern tropic in the Central Pacific, where the convergence takes place between the ENE trade wind, fed by the tropical air from the Easter Island anticyclone, and the ESE trade wind, fed by the polar air driven by the high-pressure belts in the region of the Kermadec Islands. This is the South Pacific Convergence Zone (SPCZ) (G. Cauchard and J. Inchauspe, 1976);

A region where the divergence is positive between these two convergence zones, covering the whole eastern Pacific from 30°S to 5°N and following the equator as far as 120°E.

The vertical circulation in the troposphere (rising in convergence belts and subsiding in divergence areas) is directly linked with the rain-cloud formations.

W i n d   s t r e s s   c u r l

This factor is important in calculating ocean transport in circulation models and also in determining the topography of the thermocline in the equatorial current system, since wind vorticity is directly linked to the divergence of surface currents, and thus to vertical ocean circulation.

It is calculated on the basis of the components τx and τy (counted positively eastwards and northwards respectively) of the frictional drag using the formula:

$$\text{curl } \tau = \frac{\delta \tau y}{\delta x} - \frac{\delta \tau x}{\delta y}$$

and its average annual distribution is illustrated in Figure 5B in which the isolines represent intervals of $4 \times 10^{-8} N.m^{-3}$.

In the Northern Hemisphere, the North-easterly trade winds region is divided into two zonal belts, one with negative curl, the other with positive curl, to the North and South of 15°N, respectively. Their maximum positive values are found in the Central Pacific, in the Inter-Tropical Convergence Zone, and at the northern edge of the doldrum zone in the western Pacific.

In the Southern Hemisphere, the eastern Pacific South-easterly trade winds region is also divided into two belts, one with negative values and the other with positive values, to the North and South of 18°S respectively. In the western Pacific, maximum negative curl values extend from New Guinea to the Society Islands, which corresponds to the South Pacific Convergence Zone.

It should be clearly understood that the divergence of surface waters and the upwelling of subsurface water are associated with the positive values of wind curl in the Northern Hemisphere and negative values in the Southern Hemisphere.

None of the annual average distributions illustrated in Figure 5 gives a very realistic idea of the actual substantial variations in terms of time and space, which can be found in the bimonthly illustrations in the reference publication (Wyrtki and Meyers, 1974).

Cloud cover

The rising vertical circulation in the ITCZ and SPCZ maximum convergence belts carries up the warm and humid air from the lower levels. This air cools as a result of adiabatic expansion and its initial water-vapour content condenses in the form of fine visible droplets which remain in suspension and consitute cloud formations. In certain conditions, they may become larger and fall as precipitation.

Satellites are a revolutionary tool for observing the distribution of these ephemeral cloud formations over the vast expanse of the ocean. Since the launching of TIROS I in 1960, cloud analyses have provided information of interest to environmental specialists and, from TIROS IX (1965), a view of the cloud structure over the whole surface of the Earth was obtained.

In 1976, Sadler and others used the nebulosity code categories of the National Environmental Satellite Service (NESS) of the National Oceanic and Atmospheric Administration (NOAA), US Department of Commerce, Washington D.C., USA, for the 1966-1973 period (satellites TIROS IX and X; ESSA 1,3,5 and 7; ITOS 1; NOAA 1 and 2). Using a table of equivalence between these categories and cloud cover expressed in "octas", they were able to integrate values for each mesh of a 2.5° (latitude and longitude) grid. The end product is presented in the form of monthly cloud-cover maps for the Pacific Ocean from 30°S to 60°N. The maps for February and August are reproduced in Fig. 6.

Clouds are not distributed at random, but are organized into highly coherent formations linked with low-level winds and ocean temperature. Their persistence throughout the year and their very limited seasonal movement reveal how space is organized into clearly defined minimum cloud-cover and maximum cloud-cover geographical areas.

Maximum cloud cover areas

In the 30°N to 30°S belt, the noticeable features are:

The stratiform accumulations in the eastern Pacific off the Californian and Peruvian coasts. These are "low" clouds produced by the trade wind inversion that occurs at the eastern edge of the anticyclonic cells of Hawaii and Easter Island. They comprise stratus and strato-cumulus formations which are not rain-bearing;

A near equatorial transoceanic zonal belt between 5°N and 10°N. This consists mainly of cumulo-nimbus - clouds of considerable vertical dimensions generating heavy precipitation - linked with the ITCZ. At the extremities of this belt, variations occur as a result of continental influences:

East of 120°W, it expands from a minimum of approximately 5°N from December to March, to a maximum of approximately 10°N from June to August;

West of 150°E, it is found from June to October when the low-level flow is an extension of the Asian summer monsoon, and varies in phase with this monsoon. It expands from a narrow belt extending from 5°N to 10°N in June to as far as 20°N in July and August, subsequently receding back to 5°N by October;

In the central portion, between 150°E and 120°W, it is very pronounced throughout the year between 5°N and 10°N (between 125°W and 140°W, cloud cover exceeds five octas for every month except April);

A belt in the South-west Pacific consisting of cumulo-nimbus generated by deep convection. It extends throughout the year from New Guinea to the Solomon Islands, with values of more than 4.5 octas and areas of five-octas over relief features (orographic effect). During the Southern Hemisphere summer, it intensifies and spreads South-eastwards in association with the Asian winter monsoon (December - February) and can reach as far as the Society Islands, where it links up in the South with an area of maximum cloud cover resulting from an accumulation of stratiforms on the eastern edge of the sub-tropical high pressure zones of the Kermadec Islands. This belt is linked with the SPCZ.

Minimum cloud cover areas

These are areas where the air is subsident and thus precludes virtually all precipitation. This is the case in:

The Australian continent in the southern winter, where, from July to September, there is less than two-octas cloud cover between 25°S and 5°S;

The sub-tropical high-pressure belts where the salinity of surface water is at its highest;

The equatorial belt cooled by upwelling centred on 2°S between 130°W and 160°W, which corresponds to the subsident branch of the "Walker cell"; the extension westwards of this minimum cloud cover is clearly discernible in August, September and October, when values of less than four-octas reach 145°E, in association with the régime of easterly winds. The Central Pacific arid zone corresponds to this cooled equatorial belt, where the air is subsident.

Precipitation at sea

Direct measurements of rainfall at sea over a given period are very difficult to make from vessels because of the instability of the platform, spray interference and turbulence caused by superstructures. Consequently, they are extremely seldom undertaken (Reed and Elliot, 1977), and it is necessary to rely on indirect methods in order to estimate this parameter.

Extrapolation of rainfall on land

The simplest and longest standing method is to extrapolate from measurements made by the direct method at weather observation stations on land in bordering coastal regions, or, better still, on islands. Taking advantage of the high number of islands in the Central and western Pacific, Taylor, 1973, published monthly rainfall distribution charts for the inter-tropical Pacific, based on observations from 200 selected weather stations. The charts for February and August, and for the annual average are reproduced in Figure 7.

Two maximum precipitation zones are identifiable throughout the year:

One East-west zone between 5°N and 10°N;
A belt stretching from the Solomons to Pitcairn.

Separating these two zones is the equatorial arid zone. Low precipitation zones are also identifiable within the subtropical high-pressure belts (subsident air).

Estimates based on weather observations at sea

This is the Tucker method (1961). It entails assigning given quantities of precipitation to some 100 weather conditions taken from the International Synoptic Code List and calculating the sum of these quantities on the basis of observations by vessels over a given area and during a given period. Dorman and Bourke (1979), prepared quarterly rainfall maps for the Pacific between 30°S and 60°N by applying this method to areas extending over 2° latitude and 5° longitude, with a correction to allow for local air temperature.

The two maximum precipitation belts corresponding to the convergence zones (ITCZ and SPCZ) are clearly shown, as is the extension of the arid equatorial zone to 180°. In the Southern Hemisphere, the heavy rainfall in the western portion contrasts with the relative aridity of the eastern portion.

The March-April-May quarter is the least rainy throughout the inter-tropical zone. In the Southern Hemisphere, the December-January-February quarter is the wettest, while, in the Northern Hemisphere, between the equator and the tropic, the June-July-August and September-October-November quarters are those with the heaviest rainfall.

Estimates based on satellite observations

Analyses of cloud cover are routinely carried out on the basis of satellite observations in various channels.

The judicious use of cloud cover observed in detail by satellite can be of great assistance in determining the major climatological features and relative rainfalls over the tropical and equatorial regions, where they are mainly of convective origin (Taylor, 1973).

A correlation between monthly rainfall measured on atolls and cloud cover over the corresponding 2.5° area was calculated by Sadler et al. (1976). The resulting coefficient of 0.54 is highly significant at the 1 per cent level. This is an excellent result in view of the fact that clouds are observed at a specific instant and averaged over 57,000 km^2, whereas rainfall due mainly to convection within cumulus clouds, is measured at one point only, once every 24 hours.

Precipitation in the SPCZ

The maps reproduced in Figure 7 (from Taylor, 1973) illustrate a number of climatic facts:

Throughout the year, the abundance of the rainfall in the western area (Micronesia and Melanesia) contrasts with the relative lack of rainfall in the eastern portion (Polynesia) which becomes increasingly pronounced towards the East. The equatorial arid belt extends to 180° and thus directly affects the Line and Gilbert Islands.

The rainfall pattern is quite normal, with the heaviest precipitation during the summer in each hemisphere and minimum precipitation in winter. This cycle is not very pronounced in the western Pacific between 15°S and 10°N, except in the area West of 135°E, where the alternation of the monsoons is a major factor. The cycle is more pronounced near the tropics (Marianas, New Caledonia, Fiji); in the Central Pacific, the strengthening of the sub-tropical high-pressure belts in winter results in a very pronounced minimum rainfall, except in a zonal belt centred on 10°S, where there is a relative maximum in June (e.g. the Marquesas). The aridity of the Central Pacific equatorial belt is typified in particular by the lack of rainfall from August to December, which is connected with the force of the South-easterly trade winds and the intensity of upwelling.

Year-to-year variability

The picture given thus far of the geographical distribution and timing of precipitation-producing meteorological factors may give an impression of immutability in the climatic features of the inter-tropical Pacific regions, particularly since seasonal variations are not very pronounced in the equatorial belt, with the exception of the western portion, which is subject to the alternation of monsoons. However, in reality, the situation is quite different, since yearly variations are very substantial. In some years, the sub-tropical high pressure of the Easter Island anticyclone collapses, the South-easterly trade winds moderate, equatorial upwelling ceases, the temperature of the eastern Pacific rises, the Walker circulation shifts eastwards, the ITCZ crosses the equator, the SPCZ moves eastwards, low-pressure areas moves into the Central Pacific, and there is torrential rain in the equatorial "arid zone", while exceptional drought conditions prevail in Australia and Indonesia and Peru is deluged.

All these factors in the interaction between ocean and atmosphere are taken into account in research into the "hydroclimate", which we shall discuss after describing the characteristics of the surface layer of the ocean environment.

OCEANOLOGY

The geographical description given of the SPC Area shows that it is essentially a large portion of the inter-tropical Pacific. How, then, should we set about describing the ocean environment?

A detailed presentation of existing knowledge would be an encyclopaedic undertaking which is beyond the scope of this paper. An explanation of the results in oceanographic jargon would be indigestible to the uninitiated; and the preparation of a bibliography of scientific writings on the area, as defined in strict topographical terms, without having direct access to Japanese and Russian publications would favour unfairly the role of other nationalities.

We have chosen, therefore, the attempt to answer a number of simple questions, namely:

What type of water can one expect to find at a given site from the surface to the ocean floor?

Where does this water come from?

What influence do the physical and chemical characteristics of the surface layer exert on the biological and atmospheric environment?

The ocean is a continuous and moving environment, without frontiers and never at rest. Its mobility is organized in a coherent fashion within a stratyfied structure, so that it is possible to follow a body of water in its trajectory through contiguous regions.

The surface waters move rapidly under the influence of atmospheric circulation and thus were recently to be found in neighbouring regions, where specific meteorological conditions may have left their mark on them.

Deep water moves more slowly; the water found in the inter-tropical belt originates in distant regions and, by the time it reaches the point of observation, it may have undergone changes in its long journey across other regions where special phenomena occur.

Consequently, the ocean environment of the SPC Area cannot be considered in isolation. The surface water will be described as a part of the Pacific inter-tropical belt and the deep water will be considered in the context of the Pacific basin as a whole.

We shall begin by explaining a few oceanographic ideas concerning data-gathering at sea and the physical processes which drive ocean circulation.

We shall then deal with the hydrology of surface water, intermediate water and deep water.

Thirdly, we shall consider information relating in particular to the SPC Area, such as the level of the ocean, equatorial upwelling and the movement of equatorial currents.

Finally, we shall consider two ways in which the characteristics of surface water in the SPC area play a dominant role, namely:

productivity related to nutrient salt content;
hydroclimate linked with the high heat content of the western portion.

Oceanographic concepts

Data gathering

The execution of oceanographic stations remains the basic procedure for the collection of data on the characteristics of the ocean environment. The resources necessary for the success of this operation are:

- A vessel suited to sea conditions encountered in travelling to the geographic location of the station (length, displacement, autonomy, sea-going characteristics, means of navigation) and to the work to be carried out (stability, manoeuvrability, lack of vibration, functional laboratories, etc.);

- Winches with the thousands of metres of flexible, corrosion-resistant, non-contaminating steel wire ropes needed to lower and raise equipment for sampling and measurement at the desired depths or on the bottom, while under way or on station;

- The equipment itself (samplers, nets, dredges, drag nets, pumps, measuring apparatus, etc.).

Hydrological station

This can be defined as the series of operations culminating in the in-place measurement of temperature and the collection of seawater samples (for subsequent physical and chemical analyses) at precise depths between the surface and the bottom at a specific geographical point in the ocean.

In carrying out these operations, a winch and crane are used to pay out, at a rate of 100-200 metres per minute, several thousand metres of steel cable four to five mm in diameter pulled by a 25 kg to 50 kg weight, after passing through a pulley which records the length lowered and a device for damping and taking up slack. The paying out of the cable is interrupted to enable sampling bottles in the open position and fitted with thermometers to be clamped to it at specified intervals. The bottles are long, hollow cylinders held in a vertical position through which the water flows freely as long as the movable valves do not close the openings at either end. When the last bottle clamped to the cable is immersed just below the surface (approximately 1 metre), the "cast" of bottles resembles an immense necklace suspended under the vessel, which is manoeuvred to counteract the action of wind and currents and to keep the cable as vertical as possible. After a few minutes delay to permit proper thermal adjustment of the thermometers with in situ temperatures at the lower depths, a "messenger" weight weighing 300 - 500 grams is released. This is a streamlined weight which can slide easily along the cable to which it is attached and which reaches a maximum underwater speed in free fall of between 100 and 200 metres per minute, depending on the shape and mass of the messenger, the state of the cable and the angle of inclination. The messenger strikes the mechanical device of the first bottle, which triggers the closing of the valves, inverts the thermometers and releases another messenger, held until then under the bottle, which in turn descends along the cable and strikes the triggering device of the next lower bottle. In this way, from one bottle to the next beginning at the surface, water samples and temperature measurements are taken at each preselected depth. When raised, the closed bottles are removed from the cable and placed on a rack in a wet laboratory where water samples for the various physical and chemical analyses can be drawn off and the thermometers read under proper conditions. The analyses are often carried out immediately on board in appropriately equipped dry laboratories. Some may be postponed and conducted on land using stored samples. The most common analyses are to detect dissolved gases (oxygen, carbon dioxide, nitrogen) pH, turbidity and dissolved nutrient salts (phosphorous, nitrogen compounds, silicon).

The duration of a hydrological station naturally depends on the depth reached and on the number of samples taken. For example, a station reaching 1,000 metres with 24 bottles on the cable lasts approximately one hour. However, for great depths, the time needed to lower and raise the cable becomes considerable, particularly since it is necessary also to send a messenger weight and wait until it has completed its descent.

This modus operandi for a hydrological station has remained virtually unchanged for a century, with improvements being made only to materials (rust-proof alloys, non-corrosive plastic materials, etc.). For some 10 years now, oceanographic vessels have been equipped with CTDO (conductivity, temperature, depth, oxygen) probes, with rosettes. This is a remote measuring and sampling unit suspended on a carrier-conductor cable, which can be lowered in situ. Sensors give a continuous flow of information on salinity, temperature, depth and oxygen content. Real-time processing by on-board computer permits this information to be used during the descent so that, within the ocean structure thus described, the most interesting depths can be selected for the taking of water samples by means of coupled bottles which can be closed by a signal when being raised.

Other station operations

We shall make only passing reference to the various operations calling for ocean-floor sampling (by coring, sediment buckets, nodule samplers, etc.) or the lowering of equipment near to the ocean floor. The difficulties that arise with cables are such that use is frequently made of "free" devices which are weighted, sink to the ocean floor, perform the sampling or measurement, release the weight (which is not recoverable) and return to the surface, where they are located and recovered by the vessel. Whether performed with or without cable, these operations are always very time-consuming at great depths.

It is worth noting, however, how information on the distribution of living organisms is obtained at sea. The main problem is in catching them, and the means used differ depending on their size and agility.

Small organisms which drift with the current and are known as plankton can be sampled by the bottle method at the same time as their sea-water environment. This is the method used for phytoplankton cells (from 1 to 100μ), the mass of which can be assessed by measuring chlorophyll pigments after filtration. It may also be used in the case of small zooplankton (0.2 to 2 mm), the weight or volume of which can be measured after filtration.

For larger organisms capable of "swimming" and known as "nekton", it is necessary to use catching equipment from which they cannot escape. The most commonly used are filtering devices dragged at the end of a cable by the vessel. The smallest are known as nets and the largest as trawls. The size of the opening, the mesh of the filter net and the towing speed increase with the size and agility of the organism. Such devices are manoeuvred by means of special winches and lowering gear. Apparatus is used to check the depths prospected and the volumes filtered (sounding devices, depth meters, volume meters). The capture of the largest nekton organisms, which are often marketable species, calls for the use of professional fishing methods.

Numerous other measurements, including optical, acoustical and magnetic measurements, can be carried out at such stations using suspended or free apparatus. This calls for large crews and sophisticated equipment, and thus sizeable vessels. This has been the approach adopted for major exploration projects; for more routine operations, the tendency is towards specialization as regards the type of vessel or, at least, the type of operation using a multi-purpose vessel.

In any event, the time spent at the station reduces the distance travelled and thus the area of ocean that can be studied within a given time.

Measurement under way

Under certain conditions (military vessels engaged in operations, merchant shipping, etc.), it may be essential or advantageous not to interrupt the progress of the vessel to gather samples or take measurements at surface level. Among the many devices developed for this purpose, the most commonly used for routine operations are bathythermographs (BT) which enable the temperature to be measured as a function of depth. From 1940 to 1970, they were "mechanical" torpedo-shaped devices, which were dragged along at the end of a cable and lowered or raised with a winch. Since then, an expendable BT (XBT) has been developed; it is linked to the vessel by a fine copper wire the thickness of a human hair which transmits the information to the vessel for recording until it is broken. Some XBTs can be released from aircraft and the information is picked up by radio from a small transmitter-buoy which remains on the surface and is also expendable (AXBT, airborne XBT).

After this brief description of the manner in which an oceanographic station, and particularly a hydrological station, is operated, let us now examine how the three main parameters physically identifying a seawater sample - temperature, depth and salinity - are measured. The importance of these parameters is due to the fact that they make it possible to calculate the density of the water <u>in situ</u>, which determines the hydrological structure of the oceans and, to a large extent, their circulation.

Temperature

The temperature at a point in the ocean and at an approximately selected depth is obtained with a special mercury thermometer which is lowered <u>in situ</u> inside a frame attached to the sampling bottles. Using this "protected" thermometer, which is isolated from pressure-induced compression by a glass envelope, it is possible to measure the volume of mercury occupying <u>in situ</u> the reservoir which is of a known capacity and has been isolated by breaking the mercury column at a specific point following the unit's inversion triggered by the arrival of the messenger. This volume is a function of the <u>in situ</u> temperature and the differing heat-expansion coefficients of mercury and glass. It is determined by reading a graduated column after the raising of the thermometer. The degree of accuracy achieved with a well made and properly calibrated thermometer of this type is 0.01°C.

Depth

The depth of the water sample and the <u>in situ</u> measurement of temperature is calculated from the readings given by another thermometer lowered simultaneously with the first thermometer, to which it is similar in construction and operation, with the exception that it is unprotected, i.e. exposed to the effects of hydrostatic pressure. The compression of its reservoir causes it to indicate an apparent temperature higher than the <u>in situ</u> temperature, the difference being a function of the depth, which can thus be determined accurately to within ± 5 m at depths of up to 1,000 m and ± 0.5 per cent at greater depths.

Salinity

Seawater is a complex solution of various salts. The main ions are found in relatively consistent proportions to one another, but the total quantity of dissolved salts varies from one place to another in the oceans, the average being 34.7 grams per litre.

Although all known elements are believed to be present in solution in seawater, six of them, known as major constituents, account for 99.3 per cent of the total. These are chlorine, sodium, magnesium, sulphur (in the form of sulphates), calcium and potassium. The remainder consists of minor constituents such as bromine, carbon, strontium, borium, silica and fluorine, at concentrations of more than one part per million, and trace elements such as nitrogen, phosphorous, iodine or iron at lower concentrations.

The accepted term used to describe this state of known solids in solution is "salinity", reflecting the fact that kitchen salt, Na Cl, represents the main constituent (85 per cent of the total dissolved quantity).

Salinity is defined as the total quantity of solid matter dissolved in a kilogram of seawater. Naturally, there is no question of determining the quantities of all constituents; as the relative proportions remain constant, the amount of chlorine only, as the main constituent, is determined. Consequently, the "chlorinity" is measured by silver nitrate titration and the salinity calculated:

$$\text{Salinity ‰} = 1.80655 \times \text{chlorinity ‰}$$

At least, this was for a long time the established procedure until technology made it possible (about 1960) to measure directly the electric conductivity of this saline solution using a salinometer which determined salinity to within less than 0.01 ‰.

Density of seawater

This is defined as mass per unit of volume (g/cm^3). This gives a quantity, $\rho s,t,p$ which is a function of salinity, temperature and pressure. It increases when pressure and salinity rise and temperature falls, according to relationships which have been calculated and the results of which are published in the form of tables, such as the "Tables for seawater density", US Naval Hydrographic Office, Pub. 615 (1952); $\rho s,t,p$ is the in situ density and may vary from 1.022 to 1.028. A quantity o, which is easier to express, has been introduced:

$\sigma = (\rho\ s,t,p - 1) \times 10^3$ and, in particular, $\sigma t = (\rho\ s,t,o - 1) \times 10^3$, giving values to atmospheric pressure expressed in g/l (grams per litre).

The reciprocal of the density is the specific volume in situ

$$\alpha s,t,p = \frac{1}{\rho s,t,p}$$

It can be broken down into two components $\alpha s,t,p = \alpha\, 35,o,p + \delta$, where δ is an anomaly of specific volume in relation to an ocean standard salinity of 35 ‰ at 0°C. A value close to δ, δT, is used which is a direct function of σt and is expressed as cl/t (centilitres per tonne).

Thus, if temperature in situ, salinity and depth of a seawater sample are known, it is possible to calculate its density and to express it in terms of its ot as g/l or in terms of its δT as cl/t.

Table II gives the equivalences between these two expressions for selected values.

Isentropic analysis

In the ocean, all water samples having the same density in situ (or the same σt, or the same thermosteric anomaly δT) may be considered as identifying an isanosteric surface on which movements of the various masses of water and their mixing take place easily with a minimum of exchange of potential energy or entropy. Such surfaces are known as "quasi isentropic".

On an isanosteric surface, the geostrophic flow can be calculated from the potential acceleration gradient in relation to a deeper reference surface (Montgomery, 1937, 1938), and by following the distributions of the characteristic parameters of a mass of water, it is possible to obtain an idea of the way in which it moves from its source and the changes or mixing which it undergoes during that movement.

On an isanosteric surface, temperature and salinity are connected variables. Table II shows for selected surfaces the temperature values corresponding to the isolines of salinity distribution, which are frequently represented, since masses of water are identified by the extremes of this parameter.

Table II

S ‰	TEMPERATURE° C						
	80	125	160	200	300	400	δT cl/t
	27,28	26,81	26,44	26,02	24,97	23,92	σt g/l
34,0	0,72	5,82			16,1		
34,2	2,86	7,02	9,5	11,8	16,8		
34,3	3,71	7,58	9,9			21,2	
34,4	4,47	8,14	10,4	12,6	17,5	21,5	
34,5	5,16	8,62	10,8				
34,6		9,11	11,2	13,4	18,1	22,1	
34,7		9,52	11,7				
34,8		10,05	12,1	14,2	18,7	22,6	
34,9			12,5		19,0		
35,0			12,9	14,9	19,3	23,1	
35,2				15,6	19,9	23,7	
35,4					20,5	24,2	
35,6					21,0	24,7	
35,8					21,6	25,2	
36,0					22,1	25,7	
36,2					22,6	26,1	
36,4					23,2		

Oceanic circulation

Oceanic circulation depends on the currents represented by moving water masses. This movement may be due to the entrainment force of the wind (drift currents) or to readjustments in the density field (thermohaline circulation).

Horizontal circulation

E k m a n   s p i r a l   (Fig. 8A)

By surface friction, the wind exerts an "entrainment force" which initiates movement of the surface layer at a velocity some per cent below its own, in a direction 45° [to the right (or left) in the Northern (Southern) Hemisphere] of the wind direction - this deviation being an effect of the Coriolis force due to the Earth's rotation; by viscosity, the first layer entrains the subjacent layer, at a lesser velocity and with an additional deviation; thus, the movement is transmitted downwards, from layer to layer to a depth D of about 100 m (50-200 m), where it becomes negligible; the transport of water embodied in the entire thickness of the layer between the surface and this depth is at right angles to the wind direction. This finding, obtained theoretically by the physicist Ekman, is fairly well substantiated in the open sea, in deep water, in an ocean without too pronounced stratification (no discontinuity layer very close to the surface); in most cases, the deviations observed are below the theoretical values, particularly in coastal regions, where the direction of the current may be very close to that of the wind and its rotation with increasing depth barely perceptible.

Oceanic gyre - Geostrophic flow (Fig. 8B)

The prevailing South-easterly or North-easterly trade winds generate East-west equatorial currents which, on encountering the continents, are deflected in the direction of the poles, to join, between 40° and 60° latitude, the circulating system generated by the westerlies; the latter cause the surface waters to drift eastwards where they are deflected towards the equator, along the continents and feed the equatorial currents; the ocean basins thus contain a vast closed anticyclonic circulating system for each hemisphere which is termed "gyre"; the Earth's rotation causes the water to move towards the centre of the gyre where its accumulation raises the level by forming an immense gently-sloping "hill" (the apex may be one metre higher than the periphery in a gyre whose largest dimension is several thousand km); the water accumulates until the force of gravity, which tends to make the water particles descend by the steepest slope, is exactly offset by the Coriolis force which makes them rise; at this stage, the movement without acceleration of the particles, follows "hillside" trajectories along contour lines perpendicular to the slope, thereby producing a "geostrophic" current.

The "hill" formed inside a "gyre" does not have its apex at the centre; owing to the direction of the Earth's rotation, it is displaced near the western edge where the steeper slope produces a higher geostrophic current velocity.

Dynamic topography

An indirect method of determining the velocity of currents in the surface layers is to determine the pressure field from density measurements of samples taken at hydrological stations; calculations are made of the mean densities of water columns at various stations between the surface and a deep reference level where it is assumed that there is neither current nor horizontal pressure gradient. If this assumption is correct, all the water columns of stations above the reference depth should have the same mass. However, for this to be the case, a water column with a low mean density must have a greater volume, and hence be higher than a column with a high mean density; by calculating the column height necessary at each station to obtain a zero horizontal pressure gradient at the reference depth, a topography may be determined for the ocean surface (or for any other intermediate isobaric surface); this topography is called "dynamic", for it is related to the vectors of geostrophic currents. The steeper the slope, the higher their velocity, and their direction in general is parallel to the lines of equal height in relation to the zero movement reference depth. Figure 8C presents a simple case of five stations; the surface topography shows that the apex of the "hill" should be near station 2.

Vertical circulation

In the surface layer, the horizontal wind-driven circulation may generate a local vertical circulation, which is then said to be "wind-induced".

At an entirely different scale, atmospheric conditions peculiar to certain oceanic regions affect the temperature and/or salinity of surface waters to such an extent that their sharply increased density makes them sink: this vertical circulation is termed "thermohaline".

Wind-induced circulation

When the horizontal flow of the wind-induced surface circulation is divergent, it is compensated for by vertical circulation of deep water towards the surface called "upwelling".

- Equatorial upwelling (Fig. 8D)

Due to the effect of the Earth's rotation, the westbound part of the flow of a current in the surface layer of an equatorial zonal belt in the Northern Hemisphere is deflected towards the North (to the right), while the part in the Southern Hemisphere is deflected towards the South (to the left); the surface water moving away from the Equator is replaced by an upwelling of deep water towards the surface.

- Coastal upwelling (Fig. 8E)

In ocean regions where a coastal current flows at the surface towards the Equator, along the western edge of a continent, the deflection of the flow by the Coriolis force causes the surface coastal water to flow towards the open sea and be replaced by an upwelling at the coast.

- Langmuir circulation (Fig. 8F)

The centres of large subtropical ocean gyres are high-pressure zones where the winds are generally light but stable in direction; in such conditions, large parallel narrow strips are observed at the ocean surface, more or less in the wind's path, where the flow is alternately convergent and divergent, so that circulation cells form, transverse to the current, with ascending and descending vertical branches. These cells have been given the name of Langmuir, who was the first to observe in 1938, in the Sargasso Sea, accumulations of algae along the convergence lines.

This vertical circulation induced by the local wind affects only the surface layer of the ocean to a maximum depth of 200 m (Ekman layer).

Thermohaline circulation

Vertical circulation in the oceans is the result of considerable density changes in the surface waters of certain regions with extreme climatic conditions.

On the Antarctic continental shelf, in the Weddell Sea, the temperature of the surface water drops as a result of winter cooling, and its salinity increases due to ice formation, factors which give it a very high density (-1.9°C; 34,62 ‰; σt = 27,89) the water then descends along the continental slope, mixing with the warmer Circumpolar Antarctic Water (0.5°C; 34.68 ‰; σt = 27.84) to form, towards 4,000 m the Antarctic Bottom Water (- 0.4°C; 34,66 ‰; σt = 27,86), which spreads along the bottom over the greater part of the oceans where it can be identified, mixed with North-Atlantic Bottom Water, itself formed in winter at the surface, off Greenland, where it sinks and flows southward.

In the central regions of large subtropical oceanic gyres, evaporation far exceeds precipitation, so that the surface salinity of the trapped water gradually increases, as does its density. This dense, very saline water, which is produced continuously at the surface, tends to sink and accumulate in the pycnocline, where it spreads at a level corresponding to the density acquired at the surface.

Stratification of the ocean environment

In the zonal belt between latitudes 30°N - 30°S, the energy balance at the ocean surface shows an overall surplus; that is to say, the solar energy absorbed exceeds losses due to radiation and to conduction and evaporation; the temperature of the surface layer therefore tends to rise, and the highest temperatures are, in fact, found near the equator, where the seasonal variations in solar radiation are least pronounced. However, this rise in temperature is limited by oceanic circulation which transfers heat to high latitudes with an energy deficit.

Only the surface layer is directly affected by this heating; the turbulence generated by waves and wind-driven currents then gradually mixes this warmer water with that of the subjacent layer, thereby increasing the thickness of the heated surface layer. However, the higher temperature reduces the density, and the gravitational stability of the balance of the layers becomes increasingly pronounced; turbulence, which itself decreases with depth, is thus less and less effective for propagating heat downwards.

The same applies in the intertropical zone as regards the considerable changes in salinity which may be brought about by abundant precipitation or high evaporation in warm surface water; they affect only a relatively limited "mixing layer".

An entirely different situation is obtained in high latitudes with an energy deficit; continuous cooling of the surface layer by radiation produces water of high density with a tendency to sink due to the hydrostatic instability thus created: the "cold" may thus spread downwards in the absence of any turbulence caused by external forces; consequently, the temperature varies very little from the surface to the bottom, and variations in the salinity of the surface water (reduction through precipitation, melting of ice or, on the other hand, increase from icing up) may have an appreciable effect on the density. These cold, dense waters sink and spread to low latitudes, where they reappear deep down below the warm light waters of the surface layer.

Figure 8G shows the vertical profiles of density, temperature and salinity for three oceanic regions, by latitude:

1. Equator
2. Tropics (20–25°)
3. High latitudes (50–60°)

For reasons of stability, density generally increases with depth; in the intertropical belt, a thin layer of low-density surface water is separated from the deep high-density water by a layer whose density rapidly increases with depth - the "pycnocline"; the pycnocline is barely noticeable in higher latitudes and may even be completely absent in some of these regions where vertical mixing is then possible.

A "pycnocline" is mainly the result of the rapid fall in temperature with increasing depth in a "thermocline", which is particularly marked in low latitudes; the rapid variation in salinity in a "halocline" may also affect the density; a halocline generally means a decrease in salinity with increasing depth in low latitudes, but may also correspond to increasing salinity in some high-latitude areas.

Thus the action of the thermal factor results in the very pronounced vertical structuring of the intertropical marine environment, which appears as a superposition in quasi-horizontal layers of different water masses whose density increases with depth; this density gradient lends a high degree of stability to this structure by making vertical exchanges between these different waters difficult and slow, so that they long retain, in the course of their horizontal movements, the characteristics acquired when formed at the surface.

The following water masses are identifiable in the Pacific zone 30°N - 30°S:

The waters of the upper layer formed inside this zone; this layer comprises the mixed surface layer as well as the upper part of the thermocline;

The intermediate waters formed towards 50° - 60° in each hemisphere, which occupy the lower part of the thermocline;

The deep waters formed mainly in high latitudes near the Antarctic continent: they are situated under the thermocline, as far as the bottom, and are extremely homogeneous.

Hydrology of the intertropical Pacific

Knowledge of the marine environment

Our present knowledge of the oceans is the result of a slow and patient accumulation of observations, measurements and discoveries by vessels fitted out by the principal maritime nations; it is the result of efforts by the international scientific community as a whole, despite the rival ambitions which often motivated the early expeditions; at a very early date a desire for knowledge and a need to communicate emerged which have persisted to the present day.

The vastness of the oceans calls, indeed, for the collaboration of several vessels to carry out an exploration programme; thus prestigious transoceanic expeditions like those of HMS Challenger (1872-1876) around the world, and the Meteor (1925-1927) in the Atlantic, were followed by joint international expeditions or regional surveys.

As regards the Pacific, international efforts in the oceanographic field, which developed particularly after the Second World War, were known by the name of NORPAC (USA - Japan collaboration in 1955, 20 ships, transpacific survey North of 20°N) and EQUAPAC (USA - Japan - USSR - Australia - France collaboration in 1956, 10 ships, survey of the equatorial zone 15°S - 20°N, West of 135°W).

Similarly, the Pacific Oceanic Fishery Investigations (POFI) of the US Fish and Wildlife Service, which had been studying the Central Equatorial Pacific since 1950, organized EASTROPIC in 1955 (USA - Peru collaboration, five ships, survey of the zone 20°N - 10°W to the American coast).

The oceanographic activity of each nation is exercised preferably in bordering oceanic regions; in the case of the United States, the eastern Pacific as a whole has seen cruises by vessels belonging to the US Navy Oceanographic Office, the US Coast Guard and the two major oceanographic centres, the Scripps Institution of Oceanography (based at La Jolla, California) and the Woods Hole Oceanographic Institution (based in Massachusetts); on the other side of the Pacific, Japan operates an impressive fleet fitted out by the Maritime Safety Agency, the Meterological Agency and several universities, which cruises between the Kuril Islands (50°N) and the equator; the Coral Sea and the Tasman Sea are part of the adjacent seas of Australia, where the CSIRO (Commonwealth Scientific and Industrial Research Organization of Australia) has long been active, particularly on ships of the Australian Navy; New Zealand is, of course, represented in its neighbouring oceanic zone through the activities of the New Zealand Oceanographic Institute (Wellington) and the Defence Scientific Establishment (Auckland) conducted on board vessels of the New Zealand Navy. From the French territories of the Pacific, mainly New Caledonia and, to a lesser degree French Polynesia, ORSTOM (Office for Overseas Scientific and Technical Research) organized a number of surveys from 1956 onwards, but particularly between 1966 and 1976, using vessels of the Navy and CNEXO (National Centre for Ocean Exploitation); the USSR has maintained a tradition of large-scale transoceanic surveys using vessels like the "Vityaz" or "Ob", and carried out by the Institute of Oceanology or the Academy of Sciences.

The hydrological data derived from all these surveys have been stored in "banks" at the world level (World Data Center A and World Data Center B), or at the national level (National Oceanographic Data Centre, Washington, USA, National Oceanographic Data Bank, Brest, France, for example) where they are freely available.

These data formed the basis for establishing the charts and profiles presented by Reid (1965) and Tsuchiya (1968), which are used to describe ocean hydrology.

As regards the SPC Area, the illustrations have been selected with a view to stressing international efforts in this field: the profile along 160°W, mainly the result of research by the US ships "H.M. Smith", "Stranger" and "S.F. Baird", inter alia, bears witness to the preponderance of that nation's activities in the Central Pacific; the profile along 137°E from the equator to 30°N of "Ryofu Maru" is a tribute to the numerous surveys by all Japanese vessels between Japan and New Guinea, to which we owe most of our knowlege of the Micronesian zone; the profile along 170°E from the equator to 20°S of the N.O. Coriolis testifies the important contribution made by France to exploration of the Tropical South Pacific from Melanesia to Polynesia (cf. annex).

Waters of the upper layer

Surface layer

Circulation

Pilot charts

Most of our knowledge of surface marine currents derives from observations made by merchant and naval vessels; the differences between their dead reckoning positions and the actual positions readjusted on the basis of geographical reference points are attributed to the action of currents. This information is collected and presented by specialized services of the major maritime nations in the form of current charts, for the use of seafarers: e.g. the Pilot charts published by the United Kingdom Air Ministry (1939), those of the Japanese Hydrographic Department (data for 1924 - 1934), those of the Koninklijk Nederlands Meteorologisch Institut (Royal Netherlands Meteorological Institute) (1949), those of the Instructions Nautiques of the Naval Hydrographic and Oceanographic Service (SHOM), Paris, or - the latest - the US Pilot Charts for the Pacific Ocean (US Defense Mapping Agency Hydrographic Center, (1966, 1974), which may be seen in Tabata (1975). These charts have two drawbacks: they do not make a very clear distinction between the effects of the wind and currents on ships' courses and they provide satisfactory coverage of only the most frequented sealanes.

Consequently, it is not surprising to find, in the US Pilot Charts, the broad characteristics of surface wind circulation which forms, in each hemisphere, and "anticyclonic gyre" entrained by a combination of trade winds and westerly winds. The anticyclonic gyre of the Northern Hemisphere is represented as being divided throughout the year into two cells centred on 25°N, one set about 150°E, the other at about 150°W. The anticyclonic gyre of the Southern Hemisphere is itself composed of a stable cell around Easter Island and of variable cells in the western part of the Subtropical High Pressure Belts, one of which, in the austral winter, is centred in general on the Kermadec Islands.

As regards the SPC Area, a reference to the Instructions Nautiques, series K, Vol. VIII and Vol. IX of SHOM, Paris, shows just how far currents in the archipelagos depend on local weather, hydrographic peculiarities (filling of lagoons by swells and tides) and relief. Without going into detail, we shall merely point out, therefore, that these Pilot Charts show essentially two westbound zonal equatorial currents - the North Equatorial Current North of 10°N, and the South Equatorial Current South of 5°N - encircling an eastbound current, the North Equatorial Counter-Current, between 5°N and 10°N. However, the current North of New Guinea and the Solomon Islands turns round and travels South-east in the northern winter.

The information on the currents described is more reliable and more detailed near coasts by virtue of the numerous visual landmarks.

Geostrophic currents

These currents are calculated by the "dynamic method" established in 1898 by the Norwegian Bjerknes in the meteorological sphere, and applied to oceanography by B. Helland, Hansen and Sandstrom in 1903. The method takes account of the distribution of densities and the Coriolis force only. The current is assumed to be continuous, horizontal and without acceleration; with these assumptions, the general hydrodynamic equations are simplified and express, for a particle of water, a balance between the horizontal component of the pressure force and the Coriolis force produced by the movement itself. Continuous movement is possible only according to the contour lines of the isobaric surface; the current velocity is proportional to the slope of the isobaric surface.

We have seen how it was possible, merely on the basis of temperature and salinity measurements, to determine the dynamic distances between the isobaric surfaces on the verticals of various hydrological stations and, consequently, the "topographies" of any isobaric surface, in relation to a reference isobaric surface - where currents are assumed to be zero - which is selected as far down as possible, for there are many reasons to believe that currents are extremely weak at great depths.

In the Pacific, the geostrophic currents have been calculated in relation to the 1,000 decibar reference surface (corresponding roughly to a depth of 990 m), as most of the oceanographic stations used do not go beyond this depth. Reid (1961), Reid and Arthur (1975) and Wyrtki (1975) have compiled charts of the surface dynamic topography (Fig. 9).

The units employed are the dynamic metre or cm-dyn, or again the Joule/kilogram; 1 J/kg = 0.1 dyn-m; the isolines are current lines and the velocities of these currents are proportional to the slopes of the topography, in a ratio which varies with latitude; velocities increase with proximity to the equator; a slope of 0.1 dyn-m for 1° latitude will give geostrophic velocities of 8, 10, 12, 18, 36 and 72 cm/sec in latitudes 50°, 40°, 30°, 20°, 10° and 5°, respectively.

As in the case of drift currents, the geostrophic circulation, too, forms around a large anticyclonic gyre for each hemisphere. The main equatorial currents appear clearly; the North Equatorial Current, the South Equatorial Current and the North Equatorial Counter-Current.

New counter-currents are described:

The South-Equatorial Counter-Current at about 10°S, North of the Solomon Islands, particularly strong in March-April (Wyrtki, 1975), which flows eastward from December to April and may reach Tahiti;

The North Subtropical Counter-Current (Wyrtki, 1975; Yoshida and Kidokoro, 1967), which flows eastward, just North of 20°N;

The South Subtropical Counter-Current, between New Zealand and New Caledonia, mainly flowing East and seemingly fed, by the East Australia Current which turns North in the Tasman Sea.

Direct measurements

Direct measurements of ocean currents are carried out by two main methods:

In the first, the water mass is "marked" by an object immersed in it which "passively" follows its movements; the path of the object is determined in relation to geographical or astronomical reference points;

In the second, measurements are made of the quantity of water "streaming" along a platform which is more or less stationary in relation to the bottom (absolute measurements) or in relation to deep layers of insignificant or known movement.

A good example of the first method is the use of the log books of fishing vessels employing "long lines". These immense (several kilometres long) drop lines floating in the surface layer are excellent markers, and a study of their drift during fishing provides the key to the surface oceanic circulation on the open seas. The discovery of the Equatorial Subsurface Current (Cromwell Current) at the beginning of the 1950s in the Equatorial Pacific is attributable to this method. From data provided by Japanese fishermen covering the period 1949-1969, Yamanaka (1973) studied the surface circulation of the West Pacific in different seasons. Figure 10 reproduces his pattern of equatorial seasonal circulation in the SPC Area West of 180°; the importance of the North Equatorial Counter-Current emerges clearly.

Examples of the second method, i.e. of measurements made by current metres from an oceanographic vessel which manoeuvres in order to remain more or less stationary in relation to the deep layers, will be given later (Fig. 27 and Fig. 30).

A current metre may be fixed to a cable and lowered to various depths where it is immobilized for the time taken to carry out the measurements or else left free to descend slowly along the cable while continuously measuring the profile of the current as a function of the depth (current-profiler).

C i r c u l a t i o n   p a t t e r n

Figure 11 shows a circulation pattern for the surface layer of the Pacific Ocean adapted from Tabata (1975). The continuous arrows indicate the direction of general surface currents identified by numbers. Broken arrows indicate subsurface currents (at the equator - the Cromwell Current running from the North of New Guinea to the Galapagos Islands; and along the American continent - the California Subsurface Current towards the North, and the Peru Subsurface Current towards the South). The surface currents shown are:

1. North Equatorial Current (NEC)
2. Kuroshio
3. North Pacific Current
4. California Current
5. Subarctic Gyre Currents
6. North Equatorial Counter-current (NECC)
7. Equatorial Current (EC)
8. South Equatorial Current (SEC)
9. East Australia Current
10. Tasman Current
11. South Pacific Current
12. Peru Current
13. Antarctic Circumpolar Current
14. South Equatorial Counter-Current (SECC)

Properties of the surface layer

A. Temperature, Salinity and Density of the Surface Layer

The properties of the ocean surface layer directly in contact with the atmosphere are subject to substantial variations, both annual and inter-annual.

M e a n   s e a s o n a l   d i s t r i b u t i o n s

To illustrate the mean winter and summer distributions for a "normal" year, we have borrowed from Reid (1969) charts established on the basis of data collected at more than 2,500 hydrographic stations, for each season between 1875 and 1968, by scores of vessels operated by the leading maritime nations (Reid, 1965).

T e m p e r a t u r e   (Figs. 12 and 13)

The SPC Area is characterized by very warm surface waters of low annual thermal amplitude. The temperature on the southern boundary descends to only 22°C in the austral winter South of New Caledonia and near Australia, and remains above 24°C in the northern winter North of the Marianas. It is above 28°C throughout the area in the Northern Hemisphere in the northern summer and above 26°C South of the equator in the southern summer. Temperatures higher than 29°C are found over almost half the SPC Area in each hemisphere during the respective summers, which given an average mixed layer thickness of around 100 m, represents the most formidable reservoir of heat stored at the surface of the planet.

Cooling takes place throughout the year in an equatorial belt East of the dateline, due mainly to equatorial upwelling and, to a lesser extent, to westward advection by the Equatorial Current of the colder waters of the Eastern Pacific.

According to Wyrtki (1965), who subjected data provided by merchant vessels between 1947 and 1960 in the Pacific North of 20°S to harmonic analysis, the amplitude of the annual variation is minimal in the warmest part of the West Pacific, corresponding roughly to the SPC Area (less than 0.5°C between 5°N and 5°S, below 1°C between 10°N and 10°S, less than 2°C between 20°N and 20°S), and the amplitude of the semi-annual variation is negligible. This low annual variability is confirmed by White, Meyers and Hasunuma (1982) for the North-west Pacific. In the SPC Area, they used the hydrological stations along 137°E occupied each winter from 1967 to 1974, between Japan and the Equator, by the R.V. "Ryofu Maru", of the Japan Meteorological Agency and numerous meridional profiles between 30°N and the Equator from 125°E to 180° made between 1968 and 1973, with mechanical bathythermographs (MBT) by ships of the Japanese Far Seas Fisheries. As White and Wylie (1977) had already shown, the annual variation from the Equator to 17°N is very slight.

S a l i n i t y (Figs. 12 and 13)

On the other hand, the surface waters of the SPC Area present sharp contrasts in salinity:

In the central part, the South Tropical Water is characterized throughout the year by maximum salinity higher than 36 ‰, which forms in the NE quadrant of the subtropical gyre where intense evaporation exceeds the moderate precipitation. This water tends to accumulate in the axis of the gyre, just East of French Polynesia (Rougerie, Marec, Picard, 1982).

In the western part, on the other hand, abundant precipitation, well in excess of evaporation, is responsible for the formation throughout the year of desalted water - the Equatorial Water. The lowest salinity, less than 34 ‰, is found in the northern summer in Micronesia. Values below 34.5 ‰ are usual in the southern summer in a zone extending from the North of New Guinea and the Solomon Islands as far as the Fiji Islands. These low salinities correspond to the convergence zones and precipitation described above in the context of meteorological features. Desalination in the West Pacific is mainly due to local rainfall (Donguy and Henin, 1976) and secondarily to advection by counter-currents (Jarrige, 1968).

Salinities higher than 35 ‰ are found in winter in the North of the SPC Area, in the region of the Marianas Islands, for this archipelago is located in the part closest to the axis of the North subtropical gyre where the North Tropical Water, - which is characterized by a maximum salinity exceeding 35,5 ‰ in this season - accumulates; this figure is reached in the low precipitation zone centred on 20°N, 170°E (cf. Fig. 7).

The influence of the South Tropical Water extends into the Northern Hemisphere. The 35 ‰ isoline crosses the equator between 140°W and 140°E; this extension is related to the equatorial circulation (advection and upwelling).

D e n s i t y (Figs. 12 and 13)

The parameter used in the charts is $\sigma t = 10^3$ (ps,t,o -1).

The lightest water (σt = 21.5) resulting from a combination of high temperature and extreme desalination is found in the oceanic regions of the intertropical belt covered by the ITCZ and the SPCZ. Along the equator, density decreases steadily from East to West from ot = 25.0 to σt = 22.0.

In the SPC Area, light water ($\sigma t < 23.0$) occupies the West part in all seasons, the lowest densities being observed in each hemisphere during the respective summers. Increasingly dense water is encountered towards the East in the Central Pacific. The seasonal variation is slight.

Inter-annual variability

Southern oscillation and El Niño

The mean distributions just described for the intertropical Pacific represent the commonest meteorological conditions, namely a situation where a powerful Easter Island anticyclone and well-defined equatorial low pressure over Indonesia produce a pronounced Walker cell; the warm, light waters are held in the West by well-established trade winds, where their salinity is diluted by abundant precipitation.

In some years, however, a different situation obtains, when the high pressure of the Easter Island anticyclone weakens or subsides, while the Indonesian low pressure rises. This alternance of the two relative pressure situations between the South Pacific and Indonesia is known as the "Southern Oscillation" (SO); its periodicity - about three years - is irregular.

In situations where the Easter Island anticyclone is relatively weak, the South-east trade winds abate and can no longer hold the warm light waters in the West where the ocean level falls. These waters tend to return eastward, where the level rises. The Walker cell is less pronounced and less extensive. The convergence zones are shifted; ITCZ closer to the equator and SPCZ towards the Central Pacific. The heavy precipitation moves from Indonesia to the Central Pacific. In Peru, the highly productive cold waters produced by the usual coastal upwelling are replaced by impoverished warm waters of equatorial origin which extend far southward, with disastrous consequences for the fauna (high mortality rate) fisheries (destruction of anchovy stocks) and the national economy as a whole (torrential rains, landslides, floods ...). This phenomenon is known locally by the name of El Niño (EN) (Wyrtki, 1982). J. Berknes (1966) was the first to realize that this local Peruvian phenomenon was linked with the meteorological anomalies observed throughout the Pacific in connection with the Southern Oscillation.

Although it was long the habit to refer to El Niño years (eg 1957-1958, 1965, 1972, 1976, ...) to denote the main climatic anomalies in the Tropical Pacific, the present tendency is to study the phenomenon on a Pacific-wide, or even global scale within the context of the full complexity of ocean atmosphere exchanges, which phenomenon is known as ENSO (El Niño, Southern Oscillation).

Inter-annual variability in the SPC area

In the North-west Pacific, the observations made by the Nakai Regional Fisheries Research Laboratory on board the "Shunyo Maru" between 1958 and 1964 (Yamanaka, Anraku and Morita, 1965), those of the "Kagoshima-Maru" in 1958 for the International Geophysical Year (Takahashi, Chaen and Ueda, 1960), those of the Japan Meteorological Agency from 1967 to 1974 on board the "Ryofu Maru" along 137°E (Masuzawa and Nagasaka, 1975), those of the Japanese Far Seas Fisheries from 1968 to 1973, together with the observations of the Japanese ore ships between Japan and New Caledonia from 1969 to 1978 (C. Henin and J.R. Donguy, 1979), provide material for describing annual situations more or less conforming to "normal" or widely diverging therefrom.

In the South-west and Central Pacific, the surveys carried out by the Noumea ORSTOM Centre since 1956, and particulary the collecting network observations made by merchant vessels from Noumea, made it increasingly easy to follow the temperature and salinity variations over a given period. Half-yearly salinity charts were succeeded by quarterly charts and, in 1975, by monthly charts (J.R. Donguy, C. Henin, 1978, 1976).

All the analyses establish a very sharp inter-annual variability, depending on the weather, and particularly striking anomalies in 1957-1958, 1965-1966, 1972-1973 and 1976-1977, which were El Niño years.

In particular, the maximum thermal content normally encountered West of 170°E moves towards the South Central Pacific after an El Niño; this shift is presumably attributable to variations in the depth of the mixed layer which diminishes in the West, to an increase in the temperature at the centre due to the cessation of equatorial upwelling and to the advection of warm waters from the West towards the ocean centre (C. Henin and J.R. Donguy, 1980), (Donguy et al., 1982)

Sharp anomalies also occur in the salinity distribution in El Niño years. The advent of winds from the western quadrant West of 160°W produces a convergence at the equator of desalinated waters which replace the saline water of the normal equatorial upwelling. The shift in precipitation towards the Central Pacific cancels out the normal desalination North of the Solomon Islands and Vanuatu, which are invaded by the tropical saline water entrained by the South-easterlies driven by the still relatively strong Australian anticyclone (J.R. Donguy and C. Henin, 1976).

B. Dissolved oxygen and nutrient salts in the surface layer

All the gas components of the air are found in solution in the ocean layer in contact with the atmosphere. At a first estimate, they are dissolved in quantities such that the corresponding partial pressure is in equilibrium with their partial pressure in the air; ocean-atmosphere exchanges may occur in the direction necessary to restore this equilibrium.

Of these gases, the ones most studied are oxygen and carbon dioxide, for they participate in the physiological cycles of living organisms. In photosynthesis, the dissolved carbon dioxide is used and oxygen produced. All living organisms continuously consume oxygen and emit CO_2 in respiration. The mineralization of all inert organic matter consumes oxygen and produces CO_2.

C y c l e o f o r g a n i c m a t t e r

The phenomenon of photosynthesis is the basis of primary production. It enables plants to use solar energy to synthesize their living matter from the mineral elements available in the ambient environment: carbon comes from carbon dioxide, hydrogen from water, phosphorus from a phosphate, nitrogen from its various oxidized forms, chiefly nitrate, silicon from a silicate; synthesis is effected through a complex organic molecule - chlorophyll - which is specific to the plant kingdom.

In the ocean, the essential factor in primary production is the presence of phytoplankton, composed of microscopic algae which utilize nutrients (nutritive mineral salts) in the surface layer where solar energy is sufficient (euphotic layer); they are autotrophic. The fate of living plant matter thus synthesized is to be consumed by herbivores (secondary production) which themselves may serve as food for carnivores (tertiary production), etc. In this way, it enters a complex network known as the "trophic web", composed of heterotrophic organisms, that is to say, organisms whose metabolism depends on the use of pre-synthesized organic matter.

Organic matter in the ocean is found in three forms: living (living organisms, from whales to bacteria), detritus (dead remains or the remains of organisms excreted after partial digestion) and dissolved (particles less than 0.45μ).

The utilization of organic matter by heterotrophic organisms leads to a progressive degradation of the molecules and finally, at the end of the process, to the most oxidized forms of the constituent elements: the water, carbon dioxide and nutrient salts used during photosynthesis.

This process of mineralization of organic matter is very largely the result of bacterial activity and culminates in the restoration of the elements to the environment in ratios which are fairly constant in relation to the oxygen consumed (atom ratios 276/106/16/1 for O/C/N/P).

Photosynthesis is restricted to the euphotic surface layer, where the light energy is sufficient.

Respiration takes place at all depths except for anoxic waters, where the higher living organisms cannot live.

The mineralization of detrital organic matter which takes place during sedimentation is widely distributed from the surface to the bottom; in the euphotic layer, the products of mineralization are reused for photosynthesis, while in the intermediate and deep layers, the products accumulate.

D i s s o l v e d o x y g e n

Surface concentrations in the intertropical zone correspond in general to the maximum solubility in water in contact with standard water-vapour-saturated air at a pressure close to one atmosphere, for the local temperature and salinity.

Temperature is more important than any other parameter for determining this solubility. The concentrations in the mixing layer will vary little within the limits 34-36 ‰ and will range from 5.0 ml/l at 23°C to 4.5 ml/l at 30°C.

The mixing layer, which is more or less isothermal, contains water at all depths, which has recently been in contact with the atmosphere. Moreover, the oxygen produced by photosynthesis exceeds the oxygen consumed <u>in situ</u>. The concentrations of dissolved oxygen are therefore generally at saturation or even slightly over-saturated in the most productive waters.

Below the mixing layer and thus below the euphotic layer, there is a net consumption of oxygen and the concentrations are below the saturation corresponding to the <u>in situ</u> temperature.

In the SPC Area, the oxygen concentrations of the surface layer correspond to the foregoing description.

N u t r i e n t s

Figure 14, borrowed from Reid (1962), shows the distribution of phosphate for the Pacific, at the surface and at 100 m. As may be seen at once from a comparison with Fig. 9, this distribution is controlled by the circulation. In the North and South subtropical anticyclonic gyre, the surface layer is practically devoid of nutrients at depths exceeding 100 m. In the Intertropical Pacific, the richness of the eastern part contrasts with the lack of nutrients in the western part where surface concentrations exceeding 0.25 mmole/m^3 are found only in an equatorial belt narrowing towards the West. This relative enrichment is due to equatorial upwelling, whose efficiency diminishes West of 160°W. The West-east rise of the thermocline in the equatorial belt is reflected in the fact that concentrations at 100 m increase towards the East. Counter-currents are responsible for relative minima in the higher concentrations at 100 m in the eastern part.

Nutrients of the mixing layer in the SPC Area

The lack of nutrients in the mixing layer throughout most of the SPC Area has been attested by all the oceanographic surveys. This is illustrated:

For the Central Pacific, by the PO_4-P section along 160°W (Fig. 16);

For the West Pacific: in the North, by sections PO_4-P, NO_3-N and Si. along 137°E (Fig. 32) (Yoshimura, 1972, Masuzawa <u>et al.</u>, 1970), and in the South, by the sections along 170°E (Fig. 29) (Rotschi, Hisard, Jarrige, 1972).

At the surface and over an average thickness of 100 m we may expect to find, throughout the SPC Area, mmole/m^3 concentrations below 0.25 for phosphate, 0.5 for nitrate and 2.0 for silicate. In a narrow 3°N-3°S equatorial belt, equatorial upwelling may enrich the surface water (e.g. at 170°E: 0.40, 2.0 and 2.5 for P, N and Si, respectively).

Various authors have drawn attention to this depletion of nutrient salts (Rotschi and Wauthy, 1973; Donguy, Henin, Jarrige, Rougerie, 1974).

Variability

Outside coastal waters, where continental silt may lead to temporary local enrichment, equatorial upwelling is the most important mechanism for bringing richer subsurface water to the surface. The variability of equatorial enrichment is bound up with that of the local winds. In the case of winds from the eastern sector between April and October, upwelling is present West of 180° and particularly pronounced in June, July and August (Fig. 28 and 29), while in the case of light winds or winds from the western sector (North-west monsoon) from November to March, the warm, poor waters with low salinity are found in a thick layer (100 m) at the equator itself (Figs. 31 and 32).

Waters of the upper thermocline

The distribution of water masses in the upper thermocline of the SPC Area is illustrated by charts borrowed from Tsuchiya (1968), which cover the Pacific intertropical belt 20°N-20°S. These charts show the distributions of depth, acceleration potential, salinity and oxygen concentration for four surfaces of constant thermosteric anomaly, as well as the distribution of the layer thickness corresponding to 5 cl/t above and below the nominal value.

The selected surfaces are 160, 200, 300 and 400 cl/t. Their positions in the thermocline may be seen in the sections contained in Figs. 15 and 28; 625 oceanographic stations operated between 1952 and 1961 by the international community in various seasons have been used (Figs. 17, 18, 19 and 20).

D e p t h

The absence of a 400 cl/t surface at the surface in the East of the SPC Area South of the equator clearly shows the contrast in the Southern Hemisphere between the light waters of the West Pacific and the dense waters of the eastern part. The thickness of these light surface waters is about 100 m.

For all the surfaces, the growing depth from East to West in the equatorial belt reflects the eastward rise of the thermocline.

The increasing depth of the 300 cl/t, 200 cl/t and 160 cl/t surfaces from the equator to the tropics in the western part indicates the sinking of the thermocline within the subtropical gyres.

In the equatorial belt, the topography of each surface shows alternating crest lines (minimum depths) and thalwegs (maximum depths) due to readjustments of the density field as a function of the juxtaposed currents and counter-currents of the equatorial system.

G e o s t r o p h i c   f l o w

The distribution of the acceleration potential calculated in relation to 500 db (1,000 db for the 160 cl/t surface) clearly shows the equatorial currents.

The most obvious are the westbound currents of the equatorial margins of subtropical gyres:

The North Equatorial Current (NEC) is present on all surfaces West of 120°W; its southern boundary, close to 10°N at 120°W, approaches the equator in the western part and is found towards 5°N North of New Guinea;

South of the equator, the main flow sets to the West on all surfaces; here is the domain of the South Equatorial Current;

Counter-currents are identifiable by eastward flows in the belt 10°N - 10°S;

The North Equatorial Counter-Current (NECC) is the most obvious. It is found on all surfaces. Bounded to the North by the NEC, it flows all the way across the Pacific from Halmahera to the Gulf of Panama;

The South Equatorial Counter-Current (SECC) is less obvious. It chiefly assumes the form of small eddy circulations towards 10°S in the West Pacific and is fairly well defined North of the Solomon Islands;

The Equatorial Undercurrent is a strictly subsurface counter-current and appears here as a narrow flow, no more than a few degrees wide, centred exactly on the equator and mainly evident in the Central Pacific between 180°W and 140°W.

Salinity

The most important water mass is undoubtedly the South Tropical Water, which has the highest salinities in the entire Pacific. It covers, at subsurface, the whole of the Southern Hemisphere in the belt 30°S - equator and its influence extends into the western part of the Northern Hemisphere.

In the area of 20°S - 130°W, East of French Polynesia, the surface water which accumulates in the axis of the South Subtropical anticyclonic gyre is subject to evaporation which exceeds the scant precipitation. The salinity increases and may reach more than 36.5 ‰. The resultant increase in density generates circulation by convection in places, whereby the denser water sinks and accumulates at the depth corresponding to its new density, thus giving rise to the continuous formation of an immense lens of highly saline, oxygen-saturated water deficient in nutrients and constituting the South Tropical Water.

On the 400 cl/t and 300 cl/t surfaces, the South Tropical Water is seen to participate, at subsurface, in the circulation of the anticyclonic gyre corresponding to the SEC. It is entrained westward and gradually approaches the equator, which it reaches North of the Solomon Islands and New Guinea.

In the central part of the North subtropical gyre, another saline water mass is formed at the surface in less extreme meteorological conditions than in the South. This is the North Tropical Water, with a salinity of up to 35.5 ‰. It sinks, accumulates at subsurface and participates in the anticyclonic circulation. Its intrusion in the NEC East of 20°N - 150°W is very plain on 300 cl/t and particularly on 400 cl/t, where salinities exceeding 35 ‰ reach the Philippines at about 10°N.

In contrast to the high salinities of tropical waters, a third important water mass is identifiable in the upper thermocline North of the equator. This is the desalinated surface water found off California and entrained by the California Current, which joins the NEC. It is clearly identified on 400 and 300 cl/t surfaces, where it constitutes the equatorial edge of the NEC, by a minimum salinity which extends westward as far as the Moluccan Passage just South of 10°N.

The NECC is fed at its western extremity mainly by the NEC (Mindanao Current). Nevertheless, the fairly extensive participation of the South Tropical Water which extends well beyond the equator on 400 cl/t, West of 180°, gives the NECC average salinities in the western part, which appear, however, as maximum salinity on the southern edge, East of 160°W. On the other isanosteric surfaces, the NECC is not marked by an extreme of salinity.

The SECC has practically no salinity identity in the thermocline, according to these data.

The Equatorial Undercurrent or Cromwell Current, on the other hand, is shown clearly on all the surfaces as having maximum salinity in a narrow belt very close to the equator, reflected by a succession of small eastward-pointing tongues in the isohalines. South Tropical Water is, in fact, injected into this subsurface current North of New Guinea and, transported eastward, appears as maximum salinity in the relatively desalinated waters of the eastern part; this continuous transport of saline water by the Cromwell Current is responsible for the maximum subsurface salinity found at 100 m and 200 m throughout the eastern intertropical zone between Mexico and Peru.

Dissolved oxygen

The extremely oxygen-deficient waters of the eastern part and the relatively oxygen-rich waters of the western part are in striking contrast.

The waters below 100 m between 20°N and 10°S along the American coasts in the Central Pacific are virtually anoxic; 200 and 300 cl/t surfaces clearly show that this very poor zone corresponds to the oceanic region not affected by the North and South subtropical gyre circulations. Due to the relative positions of the gyres and continents, the "bypassed" surface North of the equator is larger than to the South.

The oxygen in the thermocline is consumed by the respiration of organisms and particularly by the mineralization of organic particles (excreta, waste products, dead matter) which accumulate by sedimentation from the highly productive surface layer of this part of the ocean; the oxygen is removed by lateral mixing with richer, i.e. more "recent" waters, which were at the surface not so long ago.

On all the surfaces, the equatorial currents appear as oxygen maxima, particularly the NEC fed by the California Current. They are the suppliers of oxygen to the thermocline. Thus, an oxygen gradient positive from East to West is established in the equatorial zone.

The counter-currents transporting counter-gradient waters are marked by relative maxima which, at the end of their course, penetrate the eastern anoxic mass.

The NECC appears thus more or less clearly on all the isanosteric surfaces.

The SECC can be detected North of the Solomon Islands on surfaces 400 cl/t and 300 cl/t.

The Cromwell Current is perfectly identified on 200 cl/t and 160 cl/t by a narrow maximum oxygen belt at the equator stretching continuously from New Guinea to the American coasts. On 160 cl/t, it is clear that the Cromwell Current is fed through the arc of the Solomon Islands and the Bismarck Archipelago by the South Tropical Water which has crossed the Coral Sea.

Thickness of layers

This may first be interpreted in terms of thermocline intensity:

The high values found in the areas of 20°N and 20°S show that the thermocline close to subtropical gyre centres is less marked than that of the neighbouring equatorial zone where minimum values are encountered.

On 400 cl/t and 300 cl/t, the thicknesses tend to diminish from West to East in the equatorial zone; the thermocline is therefore more marked in the eastern part.

A relative maximum thickness is found on each surface in a narrow belt centred on the equator; on 400 cl/t, this belt reflects the upward spreading of the thermocline in the upper part of the velocity core of the Equatorial Subsurface Current; conversely, on 200 cl/t and 160 cl/t, West of 150°W, it shows a downward spreading of the thermocline in the lower part of that same current.

The very high values found on 400 cl/t at the Equator between 135°W and 140°W show that the mixing layer has been reached.

Layer thickness can also be interpreted in terms of the volume of water having certain characteristics. Thus the high values found in the equatorial belt East of 150°W and in the eastern part on 160 cl/t indicate the accumulation there of a considerable quantity of water with characteristics close to 12.5°C and 34.9 ‰, produced by the Cromwell Current at the end of its course (Montgomery and Stroup, 1962).

Intermediate waters (Reid, 1965)

Surface waters in the high latitudes of the Pacific are cold and their dissolved oxygen concentrations high. Precipitation exceeds evaporation and salinity is low. Consequently, these properties - low temperature, low salinity and high oxygen concentration - characterize the waters of the entire mixing layer at these high latitudes.

In view of the fact that this minimal salinity is found everywhere in middle and low latitudes at intermediate depths, the water masses possessing extreme such values have been called Intermediate Waters.

Thus, the meridional sections along 160°W (Figs. 15 and 16) show that the extension of the minimum surface salinity around 60°S in the intermediate layers of the South Pacific is situated just above the 80 cl/t isanostere; the evolution of the Intermediate Water of the South Pacific (IWSP) in its course is best followed on the surface identified by this value.

Similarly, the extension of low salinity values identified in the mixing layer at about 150 m at 50°N in the intermediate layers of the North Pacific is in accordance with 125 cl/t isanostere. The evolution of the Intermediate Water of the North Pacific (IWNP) can best be followed on the surface identified by this value.

C i r c u l a t i o n

The geostrophic currents on surfaces 125 cl/t and 80 cl/t (Figs. 22 and 24) indicate the presence at intermediate depth, of the major anticyclonic gyres resulting from wind-induced circulation in the layers nearer the surface. Their major axes shift towards the poles with increasing depth. Also, on surface 125 cl/t, are found the North and South equatorial currents, the NECC from one end of the Pacific to the other, the SECC in the western part near 10°S and between the two, straddling the Equator, a westbound current (Intermediate Equatorial Current, Rual, 1969)

S a l i n i t i e s   a n d   d e p t h s (Figs. 21, 22, 23 and 24)

Two large tongues of low salinity originating in high latitudes are entrained by this circulation at intermediate depths in the direction of the equator.

The Intermediate Water of the North Pacific

The IWNP is formed at the subsurface, at about 150-200 m, between 45 and 50°N, off the Aleutian Islands, by lateral mixing in the pycnocline with the waters of the subarctic cyclonic gyre. This lateral mixing reduces the salinity and temperature and increases the oxygen concentration of the water on 125 cl/t of the subtropical anticyclonic gyre of the North Pacific (IWNP: 33.80 ‰; 4.5°C; 4 ml/l O_2).

The IWNP thus formed is entrained in the subtropical gyre. By vertical mixing with the more saline waters of the surface layers and deep layers, its salinity increases during its journey towards the equator following the North Equatorial Current. It is identified by minimum salinity in the Western Tropical Pacific at depths varying from 300 m in the NECC to 700 m near 25°N (Fig. 22).

On 125 cl/t, a transpacific zonal belt centred on 5°S is marked by a relative maximum salinity attaining 34.75 ‰ (Fig. 21). This is thought to be the result of intense vertical mixing at the level of a subsurface counter-current.

On 80 cl/t, the extension of IWNP is clearly visible in the North subtropical gyre (Fig. 23).

The Intermediate Water of the South Pacific

The IWSP is formed at the surface near 60°S. Because of climatic conditions, its characteristics vary according to the season (ice obstruction, breaking up of drift ice, precipitation, winter, summer,....) and depending on longitude, within a range above freezing point (33,87 ‰; -1.845°C to 34.20 ‰; +3°C). This water sinks at the Antarctic Convergence (near 55°S), as is shown in Fig 24 which gives the depths, and is entrained in the anticyclonic circulation of the South Pacific. There, on 80 cl/t surface, it is possible to trace its characteristic minimum salinity which, near 55°S - 90°W penetrates the subtropical gyre (Fig. 23). Identified at increasing depths, from the surface to 1,000 m on the eastern edge, where it attains 34.40 ‰, this minimum salinity is then entrained westward. It is found again at 20°S, South of the Fiji Islands, and enters the Coral Sea (34.50 ‰; 5.2°C at 1,000 m) at the level of New Caledonia. The IWSP flow then divides into two branches (see Fig. 24 showing geostrophic currents), one towards the South along the Australian continent, and the other northwards through the arc of the Solomon Islands as far as the equator, which it reaches with the characteristics 5.3°C and 34.52 ‰ at a depth of 800 m. On surface 80 cl/t, in the mainly zonal flow of the intertropical zone, the salinity and temperature isolines run from East to West. In the zonal belt 5°S - 12°N in the West and 15°S - 20°N in the East, these parameters have relatively uniform values, close to 34.55 ‰ and 5.5.°C which characterize the minimum salinity of the equatorial water of the Pacific (Sverdrup et al., 1942).

On 125 cl/t, the progression of minimum salinity of IWSP is as easy to follow as on 80 cl/t. The same penetration of the gyre on the meridional 90°W, the same sinking in the Eastern Pacific to almost 500 m and the same westward course at this depth to enter the Coral Sea between the Solomon Islands and Vanuatu.

O x y g e n   c o n c e n t r a t i o n s   (Figs. 21 and 23)

In its source region, IWSP becomes saturated with oxygen on contact with the atmosphere, with a concentration close to 7 ml/l.

IWNP is not formed directly at the surface, but by lateral mixing with the waters of the subarctic cyclonic gyre, which leaves its mark on the subtropical anticyclonic gyre South of the Aleutian Islands. Desalination to 33.80 ‰ is accompanied by oxygen enrichment to 4.0 ml/l (Fig. 21) on 125 cl/t.

Thus, these intermediate waters start out by being rich in oxygen. But as they move about, the oxygen will be consumed by the respiration of living organisms and particularly by the mineralization of organic matter, whether carried along at the time of the original sinking or acquired on the way by the sedimentation of particles originating from surface layer production. Consequently, the longer the route of the intermediate waters, the smaller the oxygen concentration. We have seen (Figs. 22 and 24) how these waters first follow the subtropical gyres to the equator in the western part and only then reach the eastern part via an equatorial transpacific route. The minimum oxygen concentrations (less than 0.25 ml/l) are therefore found in regions reached only at the end of their journey, that is in the eastern part of the Intertropical Pacific which is "by-passed" by subtropical anticyclonic gyres and which, moreover, includes regions with the highest primary production in the euphotic layer (California coastal upwelling in the North, Costa Rica dome near the equator, Peru coastal upwelling in the South). As may be seen from Fig. 23, on 80 cl/t, the IWNP plays little part in the oxygen renewal of this layer, the main influence being that of the IWSP, which crosses the equator after passing through the Coral Sea and joins the circulation of the Northern Hemisphere along the island arcs formed by the Philippines, Formosa and Japan.

In the equatorial system, the counter-currents transport waters which are relatively richer in oxygen, in an easterly direction and their course is marked by relative maxima which penetrate the virtually anoxic eastern mass whose lowest concentrations are found throughout the layer between 125 cl/t and 80 cl/t off the Mexican coast.

C o n c e n t r a t i o n s   o f   n u t r i e n t s   (Figs. 21 and 23)

- **Phosphate concentrations**

The IWSP has a concentration of about 2 mmole/m^3 when it enters the South subtropical anticyclonic gyre, where it becomes impoverished by vertical mixing with the surface waters. The concentration tends to increase slightly towards the equator, which is crossed North of the

Solomon Islands. The intrusion of IWSP into the richer Northern Hemisphere is nevertheless marked by a relative minimum which can be traced in the northern gyre as far as 170°W. Maximum values above 3 mmole/m^3 are found in the eastern part of the hemisphere.

Taken as a whole, the phosphate distribution is the reverse of dissolved oxygen distribution.

- Nitrate concentrations

There is a very significant correlation, in a water mass, between phosphate concentrations and nitrate concentrations, both of which originate from the mineralization of organic matter, whose composition is relatively constant. The atom-to-atom ratio for N/P is thus between 12 and 15. Nitrate distributions may therefore be said to be virtually the same as for phosphate.

- Silicate concentrations

Silica is a constituent of the skeletons of certain small forms of plankton, diatoms and radiolaria. However, most organisms contain very little or none at all. The siliceous parts of these skeletons are subject to more rapid sedimentation than other organic waste matter, so that the reappearance of the silica in solution occurs at different depths and in different places from those where nitrogen and phosphorus are mineralized. A maximum silica concentration of 180 mmole/m^3 is thus found in the deeper layers, between 2,000 and 3,000 m, in the eastern part of the Pacific. In the intermediate waters, the concentrations accordingly increase with depth, and one may expect to find fairly uniform values ranging from 40 - 50 mmole/m^3 on 125 cl/t, to 120 - 140 mmole/m^3 on 80 cl/t.

Characteristics of intermediate waters in the SPC Area

In the SPC Area, the Intermediate Waters occupy the lower part of the thermocline. On the basis of the various charts presented and of the foregoing comments on nutrients, the following characteristics may be produced:

	On 125 cl/t	On 80 cl/t
T°C	7 - 10	4,15 - 5,15
S ‰	34,20 - 34,80	34,35 - 34,50
O_2 ml/l	0,5 - 4,0	1,0 - 4,0
PO_4-P mmole/m^3	1,0 - 2,5	1,5 - 3,0
NO_3-N mmole/m^3	12 - 30	18 - 36
SiO_3-Si mmole/m^3	40 - 50	120 - 140

Deep waters

These fill the entire Pacific basin underneath the Intermediate Waters which, as we have seen, occupy the lower part of the thermocline.

C i r c u l a t i o n

The figures of the Antarctic-Alaska transpacific section (Figs. 15 and 16) already show that the source of the high-oxygen content bottom water (below 3,000 - 4,000 m) has to be sought in the Antarctic, whereas lower oxygen contents are, of course, connected with conditions obtained in the Northern Hemisphere. But it is difficult to draw any major inference as regards circulation.

A map of the entire basin, showing the in situ temperature distribution at 4,500 m and taken from Knauss (1962), is much more explicit (Fig. 25). The temperature rises slowly and steadily from South to North; the coldest water enters the basin from the South and moves towards the North along its western edge, not next to Australia since the Tasman Sea is closed to the North but next to New Zealand and the Tonga-Kermadec chain.

The transpacific hydrographical sections along parallels 43°S and 28°S drawn up by the U.S.N.S. Eltanin during the Scorpio expedition in 1967, have shed fresh light on the real nature of this deep current predicted by theory.

The temperature and salinity profiles along the 28°S parallel, taken from Warren (1970) (Fig. 26), illustrate the intrusion between 3,000 m and 4,000 m along the Tonga-Kermadec ridge of saltier water (salinity greater than 34.72 ‰ with core at 34.73 ‰), moving towards the North (with isotherms sinking towards the East by a few hundred metres over a width of approximately 900 km): this water is the last vestige of the deep water of the North Atlantic transported eastwards around the Antarctic by the circumpolar current at the same time as the Antarctic bottom water (Mantyla and Reid, 1983). It can be located on the profile at 43°S, between 3,000 and 4,000 m, from 160°W to 170°W, against the Chatham ridge, by a maximum salinity of 34.73‰.

At 28°S, using these data, a geostrophic estimate of this deep current, in relation to the horizontal surface at 2,500 m which is assumed to be immobile, gives a transport of 17 Sv for the total distance from 2,500 m to the bottom (the unit of transport is the Sverdrup: 1 Sv. = 10^6 m^3 s^{-1}).

This northwards transport of deep water, characterized by maximum salinity and maximum oxygen-content, must reach the Northern Hemisphere by a fairly direct route, since traces of these extremes were found in the Marianas Trench near 12°N - 144°E (Mantyla and Reid, 1978) and on a Japan-California section along 35°N around 165°E (Kenyon, 1983) during the INDOPAC expedition carried out by the S.S. "Thomas Washington" of SCRIPPS between 1976 and 1977. The passage to the North-east of the Samoa Islands is relatively narrow (200 km at 4,000 m and 50 km at 5,000 m) and a mean current of 15 cm per second, with peaks up to 20 cm per second, bearing North-North-East, has been measured there at 4,800 m (Reid, 1969).

Characteristics of deep waters in the SPC Area

The deep waters are relatively homogeneous since they are situated below the thermocline and the extremes of the Intermediate Waters.

Between 2,000 m and the bottom, the gradients are very gradual, both horizontally and vertically. Moreover, they are virtually unvarying, the saline cold oxygenated source being deep in the Southern Hemisphere and the low-salinity source, poor in oxygen and rich in nutrients, being subsuperficial in the Northern Hemisphere.

The characteristics will therefore come within the following ranges (between 2,000 m and 6,000 m):

In situ temperature	2,5°C	-	1°C
Salinity	34,60‰	-	34,72‰
Oxygen	2,0 ml/l	-	4,0 ml/l
Phosphate	3,0 mmole/m^3	-	2,2 mmole/m^3
Nitrate	43 mmole/m^3	-	33 mmole/m^3
Silicate	180 mmole/m^3	-	140 mmole/m^3

Circulation in the SPC Area

Having given an account in qualitative terms of the masses of water to be found at various levels and of their respective movements, we should now like to turn to oceanic circulation in the SPC Area, viewed from the quantitative standpoint.

We shall deal briefly with ocean level, which enables currents in the upper layer of the ocean to be monitored. We shall dwell at somewhat more length on the measurement of currents in the equatorial belt at 150°W and 170°E; and lastly, we shall provide a short summary of equatorial circulation in the SPC Area.

Ocean level

Temperature of the surface mixing layer and atmospheric pressure are the two parameters that have most influence on the annual variation of the average ocean level. Throughout the whole of the SPC Area, the extent of the annual variation is modest, generally less than 50 mm, except in the Coral Sea and Eastern Micronesia, where it exceeds that figure owing to seasonal variations in pressure over the continents. The maximum level is reached at the end of summer in each hemisphere; on the equator, between 160°W and 170°E, the maximum level is reached in September and October when upwelling is less pronounced.

In the West Pacific, significant anomalies in the ocean level have been noted in connection with occurrences of El Niño (Wyrtki and Leslie, 1980). In a study of the level at Truk Island, Meyers (1982) showed that significant declines in level of up to 20 cm had been noted in the northern West Pacific, in phase with occurences of El Niño in the East Pacific in 1953, 1957, 1963, 1965, 1969, 1972 and 1976. The period of decline lasts for a year, from April to April, the minimum level being reached in December, in phase with the occurrence of El Niño on the coasts of Ecuador and Peru. The drop in level is combined with a rise in the thermocline following an oscillation related to the wind field.

In a comparison of the seasonal variations of the dynamic topography of the surface, relative to 500 db in the western Pacific, with the seasonal fluctuations of the ocean level measured in a dozen islands, Wyrtki (1974) showed that there was a close relationship. The real-time monitoring of differences in level between these archipelagos makes it possible to track the transports by the currents and to evaluate the advection of the thermal content of the upper layer of the tropical ocean (Wyrtki, 1978). The ocean level is currently monitored by a whole network of tide gauges installed on several islands in the Central and West Pacific (most since 1974 as part of the North Pacific experiment of the International Decade of Ocean Exploration (NORPAX), funded by the National Science Foundation).

Equatorial circulation

C e n t r a l   P a c i f i c

In the eastern part of the SPC Area equatorial circulation has been under study since 1950.

During Eastropic in 1955 (Austin 1960), sections made by Hugh M. Smith along 140°W and 155°W showed that, on the equator and around 10°N, the thermocline was nearer the surface.

On the equator, the divergence in the westward surface flow of the Equatorial Current causes an "upwelling" or rise of the subsurface water, colder and richer in nutrients, which mixes with the surface water. Temperatures lower than 24°C and PO_4-P contents higher than 1.2 mmole/m^3 have been observed at the surface (see also Fig. 15 and 16). There is an upward spreading of the uppermost part of the thermocline (above isotherm 20°, which remains virtually horizontal), this being an indication of vertical circulation in the surface layer (Fig. 27).

Near 10°N, the circulation entails a barocline structure (the isanosteric surfaces are inclined relative to the horizontal surfaces), so that, at the boundary between the NEC and the NECC, the thermocline rises and the 22°C isotherm is found at less than 50 m below the surface where the temperature, however, is 27°C and the PO_4-P content low, 0.4 mmole/m^3 or less. This crest-like structure is known as "ridging". It differs from upwelling in that there is no upward spreading of the thermocline. All the isotherms rise, but no significant cooling or lasting enrichment are discernible at the surface; vertical circulation in the surface layer is very weak.

Upwelling is related to meridional circulation (Cromwell, 1953). There is equatorial convergence at the subsurface which feeds the upwelling of water towards the surface. At the surface, the "upwelled" water moves away from the equator as far as 2° - 3°N or S, where it sinks below the warmer lateral waters along zonal "convergence" lines. The convergence line to the North is very marked and often assumes the aspect of a "front" separating waters having very different characteristics.

The most original feature of equatorial circulation is the Equatorial Undercurrent, or Cromwell Current. This is an eastward counter-current in the form of a thin strip (400 km wide, 200 m thick) found at the sub-surface, symmetrical with the equator, where velocities are greater than 25 cm per second, and may reach 150 cm per second in the central stream. Transport is of the order of 40 Sv. This counter-current is embedded in the thermocline, within the Equatorial Current flowing towards the West; it follows the rise in the thermocline from West to East. It may be caused by geostrophic flows converging towards the equator in the upper part of the thermocline, due to the pressure of the upward East-west slope of the ocean surface (Knauss, 1963).

In a quantitative study of equatorial upwelling in the Pacific, Wyrtki (1981) estimates the vertical transport due to upwelling between 100°W and 170°E to be 50 Sv., which corresponds to an average vertical velocity of the order of one meter a day. The thermal budget shows that, while the horizontal advection by the Equatorial Current does indeed help to maintain the tongue of cold equatorial water which is particularly marked from August to October, the main contribution is attributable to upwelling.

The Hawaii to Tahiti Shuttle Experiment was carried out, as part of NORPAX, in the Central Pacific from 1979 to 1980. The purpose was to study the variability in the equatorial upwelling in terms of time and space. It consisted of a group of joint operations comprising 15 hydrological sections, as well as current measurements using a profiler (Firing <u>et al.</u>, 1981) along meridians 150°, 153° and 158°W (Taft and Kovala, 1981), 35 thermal sections between Honolulu and Papeete carried out by aeroplane (AXBT) (Stroup <u>et al.</u>, 1981) and continuous recordings of temperature, salinity and current at the equator using three deep anchorages in the vicinity of 153°W (Knox and Halpern, 1981). At the same time, winds were recorded at these anchorages and at Jarvis Island (160°W on the equator). Wyrtki and Eldin (1982) report that five major upwelling incidents were identified in 18 months. Each was caused by a freshening of the trade wind for a period of 10 to 20 days and was reflected in a drop in the ocean level (of the order of 10 cm), surface cooling (from 28°C to 26°C) and a rise in isotherms of several dozen metres. In such cases, the vertical velocity at the equator is of the order of 3 m a day. When the wind abates, upwelling ceases, the thermal structure collapses and the raised isotherms again sink. Upwelling incidents are therefore regarded as the localized response to winds of limited duration. In climatic terms, equatorial cooling appears to occur over a belt 400 - 500 km wide, to be highly consistent in its longitudinal expansion over thousands of km in the Central Pacific, but to diminish rapidly towards the West as from 180°, consequent upon the increase in thickness of the mixing layer.

Current measurements using a profiler from 5°S to 5°N (Firing <u>et al.</u>, 1981) revealed a Cromwell Current centred on the equator with an average velocity core of 80 cm/second, rising to 140 cm/second near 120°. Identifiable by a velocity greater than 20 cm/second, the thickness is 200 m on average and the width from 1°30 N to 1°30 S. The transport varies from 20 to 65 Sv., the average being very close to 40 Sv (Fig. 27).

W e s t e r n   P a c i f i c

In this part of the SPC Area, equatorial circulation has been studied in particular by ORSTOM in the course of 10 expeditions (Bora and Cyclone) along 170°E, from 20°S to 4°N, between 1965 and 1968 on board N.O. Coriolis (Rotschi, Hisard and Jarrige, 1972). On each expedition, a hydrological section and measurements of dissolved oxygen, nutrients and chlorophyll were made between 20°S and 4°N. Between 4°S and 4°N direct current measurements were carried out at various depths using recording current metres with Savonius rotors (Hydro-Products) suspended from a cable kept vertical by continuous manoeuvring of the vessel against the surface current. The measurements were taken in the 0-500 m layer by comparison with the readings of a current metre fixed 1,000 m down (so that the 1,000 m - 1,500 m layer was the reference point) (Magnier et al., 1973). Figures 28, 29 and 30 give the results obtained during the Cyclone 6 expedition (18 - 27 August 1967).

- Upwelling

These expeditions showed that for most of the time there was upwelling at 170°E when the wind came from the sector East on the equator: that there was an upward and a downward spreading of the thermocline over significant thickness (from the surface to a depth of 400 m); that the rise towards the surface was accompanied by a modest cooling of the order of 1°C by comparison with the adjacent waters to the North and South; and that an enrichment in nutrients and chlorophyll in the 3°N - 3°S belt was then discernible.

With the wind from the West, there was a convergence of warm waters towards the equator where a layer of isothermal water depleted of nutrient salts and over 100 m thick could be found; the spreading of the thermocline then disappeared in its upper part.

An example of such convergence was found at 137°E in January 1970 by the Ryofu Maru (Figs. 31 and 32). No rise of isotherms towards the surface is to be observed at the equator, where warm, comparatively desalinated and nutrient-depleted water is to be found. This is normal at this longitude and in this season given the North-west monsoon.

- The Cromwell Current

The expeditions established that the Cromwell Current was in fact present at 170°E. It was found at the subsurface from 50 m to 300 m over a width of the order of 2°30N - 2°30S. It consists of two cells of current, one centred at about 100 m (30-40 cm a second), the other at 200 m (50 cm a second) separated by a layer where the flow has a strong meridional component. Deep extensions of the counter-current towards the East were present at about 3°N and 3°S. The Cromwell Current was connected to the NECC by a bridge at about 200 m at 3 - 4°N.

With a wind from the West sector, an easterly surface equatorial current was observed (Cyclone 3 expedition, April 1967) between 2°N and 2°S over a 60 m thickness. The upper cell of the Cromwell Current disappeared and was replaced by a westerly cell. The lower cell remained unchanged.

The expeditions along 170°E therefore showed that the circulation on the equator (and hence the thermal structure) in the 0 - 100 m layer depended on the wind régime.

From 1970 to 1976, the Eponite and Minepo expeditions advanced the understanding of the role of upwelling in the meridional circulation of the 160-175°E equatorial zone (Oudot and Wauthy, 1976) (Oudot et al., 1979). The validity of the circulation pattern of T. Cromwell (1953) in the Central Pacific was established for this western part. Intensity of upwelling depends upon the strength of the East-sector winds at the equator. With the wind from the South-east, the surface divergence appears slightly to the South of the equator and the "upwelled" water - cold, saline and rich in nutrients - is advected northwards as far as 3°N, where it sinks under the thick, warm and desalinated surface layer and then spreads and accumulates in the thermocline at a depth corresponding to its density. With the wind from the North-east, the divergence appears to the

North of the equator and the frontal zone to the South. With the wind from the East, a more symmetrical situation may be encountered, with divergence at the equator and sinking at about 3°N and 3°S. The upwelling is fed by a convergence at the equator of waters from the crest of the thermocline with contributions from the meridional circulation cells that enable the organic matter to be recycled. The cycle is as follows: upwelled water, photosynthesis in the euphotic layer during the advection phase, sinking along the front, storage in the thermocline and mineralization of the entrained planktonic organic matter, supply of nutrients to the upwelling. The Eponite and Minepo expeditions also showed that at about 10°S, at the southern limit of the SECC, the thermocline rose and enabled richer waters to reach the euphotic zone. Under a warm desalinated surface layer, depleted of nutrients, but of reduced thickness, an accumulation of plant biomass was to be found in the upper part of the thermocline, which is an indication of stimulated primary production. This structural feature is entirely comparable to that found at 10°N, at the northern limit of the NECC.

- Zone of formation of Cromwell Current

The study of the waters of the upper thermocline revealed that for the most part, the Cromwell Current is fed by South Tropical Water via the Coral Sea (Fig. 18). On the meridional sections along 137°E of the Ryofu Maru (Fig. 31) the entry at the equator, at about 200 m beneath the surface of oxygen-rich ($O_2 > 3,5$ ml/l) saline water ($S > 35,5‰$) from the Southern Hemisphere, and more specifically from the Coral Sea, is clearly visible. Its presence to the North of New Guinea suggests a passage through the Solomon Islands.

Transports of equatorial currents

The North Equatorial Current (velocity 5 - 15 $cm.s^{-1}$). Fed in the East by the California Current and the return flow of the North Equatorial Counter-Current, it is strengthened in the Central Pacific by a branch of the North Pacific Current. It is a broad and powerful current which covers a zonal band between 8 - 10°N and the Tropic of Cancer to the West of 180° and transports over 100 Sv. when it arrives off the Philippines (Masuzawa, 1964).

The South Equatorial Current (velocity 5 - 10 $cm.s^{-1}$). This, too, is a broad current between the Tropic of Capricorn and 10°S. Very little is known about its transport but, on the basis of a comparison with the NEC, it could reach 60 Sv.

The Equatorial Current (velocity 20-40 $cm.s^{-1}$). This flows between 5°N and 5°S and would appear to transport up to 63 Sv. westward (Montgomery and Stroup, 1962; Hisard et al., 1970; Magnier et al., 1973). It is due to the entrainment force of the East winds that come from the Southern Hemisphere and reach the equator (mainly the trade-wind of the Easter anti-cyclone). Included in this westward flow and centred on the equator and flowing towards the East there is the subsurface jet-stream which was discovered in 1952 and is known as the Cromwell Current or Equatorial Undercurrent.

West of 180°, its core lies at a depth of about 200 m. With a width of 2°N to 2°S and a thickness of over 200 m, it achieves velocities of 40 to 50 $cm.s^{-1}$ (Fig. 30). At 170°E, its transport varies from 12 to 43 Sv. (Rotschi, Hisard and Jarrige, 1972).

At 150°W, its core lies at about 120 m, where its velocity can reach 180 $cm.s^{-1}$. With 200 m thickness, the velocity may exceed 40 $cm.s^{-1}$ (Wyrtki and Kilonsky, 1982). Its transport varies from 20 to 65 Sv., estimated on the basis of profiler measurements between 5°N and 5°S (Firing et al., 1981).

In addition to this subsurface eastward transport, there are those of the two equatorial counter-currents, observable at the surface.

The North Equatorial Counter-Current (velocity 20-40 $cm.s^{-1}$). This is the more important, given that it is permanent and extends from one end of the Pacific basin to the other. It covers a relatively narrow zonal band between 3°N and 8°N and its transport varies from 12 to 61 Sv. (Montgomery and Stroup, 1962; Knauss, 1961; Wyrtki and Kendall, 1967).

The South Equatorial Counter-Current (velocity 5 - 10 $cm.s^{-1}$). This is of more limited significance and was observed for the first time only in 1959 (Reid, 1959; Wooster, 1961). Its existence was confirmed by several expeditions of the N.O. Coriolis from Noumea. It is permanently identifiable to the North of the Solomons. Its transport is very variable, ranging from 2 to 20 Sv. at 170°E (Jarrige, 1968). Its eastward extension has been studied (Donguy, Henin and Rougerie, 1976) and its transport at 140°W is believed to be 10 Sv. It is doubtful whether it exists East of 130°W, so that, in the Eastern Pacific, the SEC and EC could form a single current from 20°S to 4°N.

Figure 33 gives a simplified circulation pattern and the possible maximum transport balance in the SPC Area. The arrows indicate the direction of the flows and the numbers the quantities transported, in Sv. The probable exchanges with the inland seas of South-east Asia via the Philippines archipelago, which can be taken to be small (Soegiarto and Birowo, 1975), have been disregarded. Thus, in the SPC Area, the equatorial currents are believed to bring 220 Sv. from the East (100 NEC + 60 EC + 60 SEC). The counter-currents are believed to return 120 Sv. to the East (60 NECC + 40 Cromwell Current + 20 SECC). The difference, 100 Sv., feeds in equal parts the Kuroshio (Worthington and Kawai, 1972) via the Luzon current (Nitani, 1972) and the East Australia Current (Hamon, 1970) via a complex circulation in the Coral Sea. The Southern Hemisphere contribution to the NECC appears to be small with the Cromwell Subsurface Current transporting mainly waters from the Southern Hemisphere.

This orderly arrangement of transport balance should not conceal the fact that these currents are very variable and as yet little known, and that only the Cromwell Current has been subjected to direct measurement by current-metres, the others having been estimated by the geostrophic method. To make the position quite clear, we would point out that, prior to the discovery of the Cromwell Current in 1952, the balance shown for the equatorial system was just as satisfactory!...

Arrows without specific transport have been added to Figure 33 at about 20°-25° latitude in each hemisphere. These indicate the flows of the Sub-Tropical Counter-Currents whose existence Yoshida and Kidokoro (1967) predicted on the basis of the wind stress curl. These currents have since been observed in the Northern Hemisphere (Uda and Hasunuma, 1969) and in the Southern Hemisphere (Merle, Rotschi and Voituriez, 1969; Rotschi, 1973; Denham et al., 1981), but their permanence has still to be confirmed.

CONCLUSION: THE OCEAN ENVIRONMENT IN THE SPC AREA

We have seen above what kind of water is to be found, at all depths, in the SPC Area. As regards deep waters, the consistently extreme climatic conditions in the high-latitude regions where they are formed, the slowness with which they move, and the thickness of the layer which isolates them from surface influences in the rest of the basin, explain their homogeneity and their perenniality. Observations made decades apart have identified the same characteristics in the same region.

In the case of the waters of the upper layer in direct contact with the atmosphere, the position is entirely different. Ocean-atmosphere exchanges do, of course, take place according to known processes in geographically defined regions, but the meteorological variability is such that they are subject to anomalies in space and time that have repercussions on the characteristics of the surface water. Under the effect of oceanic inertia, these abnormal characteristics can last for a long time over relatively large areas and can be transported by the currents to relatively far-off regions where they will, in turn, influence climatic conditions. Furthermore, it is in these surface waters that man engages in most of his fishing, communication and leisure activities and which essentially constitute his immediate environment.

We have seen that, in the SPC Area, the ocean appears, at the surface, as a thick layer of warm very infertile westward-moving water. These characteristics have a direct influence on the biological and climatic environment on which we wish to comment in conclusion of the present paper.

Impoverishment of waters

The waters to be found in the surface layer of the SPC Area entered the vast noria of transpacific intertropical circulation at the eastern frontages of the two subtropical anticyclonic gyres where the coastal upwellings of California and Peru provide a degree of enrichment through injection of subsurface water (Figs. 11 and 14). A fairly significant primary production by photosynthesis then takes place in these waters at the source of the North and South Equatorial Currents. These waters will gradually grow warmer as they move from East to West and will always remain on the surface, since they are consistently the lightest in the column of water at any given place. With the tendency of this warming process to spread downwards, although with difficulty, a warm mixing layer forms and thickens, and is increasingly isolated from the deeper layers by a well-established thermocline. The surface layer very quickly becomes impoverished as the nutrient elements are consumed by photosynthesis and the removal of particulated organic matter as a result of sedimentation. Soon there remains only an increasingly small primary production which operates on the basis of the partial remineralization in situ in the isolated layer. Towards the centre of the gyres, the residence time increases and the depleted waters tend to accumulate and sink as a body. In this "subsidence" zone, no upwelling of the richer subsurface water is possible and the most extreme conditions of "oligotrophy" can be found (oligotrophy being the environment in which the nutrient concentrations are low and organic production small). This is the area of "ocean deserts", that is very clear water resulting from the dearth of suspended particles (A.P. Lisitzin, 1970) and hence very blue (deep ultramarine) water, where living organisms are few and far between. Virtually only highly migratory animals, most often in the reproductive phase, are found there.

The highly saline tropical water which, formed in the centre of a gyre, penetrates and becomes part of the circulation at the top of the thermocline, forms throughout the intertropical zone a "floor" where the density gradient is very pronounced. It is a very effective device for curbing the sedimentation of organic particles and "trapping", as it were, the nutrients produced by their mineralization. Thus, as they are transported, the tropical waters at the subsurface retrieve nutrients and carry them towards the equator where upwelling can take them to the surface for recycling by photosynthesis.

Upwelling is one of the mechanisms whereby the inherent poverty of tropical waters can be counteracted. In the SPC Area, as has been seen, it can be found at the equator. Another transoceanic extension mechanism is the rise in the thermocline at near 10°N and 10°S, at the edge of the equatorial counter-currents, which makes nutrients available to the phytoplankton in the euphotic layer (Fig. 14 gives the relative maxima of PO_4-P at 100 m in the vicinity of 10°N, 10°S). Lastly, a third mechanism for the input of nutritive elements in ocean surface waters results from the process of rock weathering/leaching by rain/collection by drainage basin, which plays a very localized role around the small, high isolated islands of Polynesia but only has any effective significance on an ocean-wide scale in the Melanesian region where highlands are numerous and rainfall is very heavy. Figure 34 reflects very clearly the predominance of physical factors (thermal balance, circulation) over the organic production of the Pacific. The two huge subtropical gyres are deserts. The 10°N - 10°S equatorial belt has a higher production which, however, decreases from East to West. The coastal waters, including those which wash the Melanesian archipelago, are more productive than those of the open sea. This distribution of production persists, of course, throughout the next links in the food chain, as is apparent from Figure 35 which gives the distribution of zooplankton netted in the 0 - 150 m layer (Reid, 1962).

Most of the hauls (1335 out of a total of 2005 hauls) were made between 1950 and 1960 with a net with a 1 m opening, 5 m long and 0.25 mm and 0.55 mm meshes, during the NORPAC, TRANSPAC, EQUAPAC and EASTROPAC expeditions. In the SPC Area, only the equatorial region, particularly to the North of New Guinea, escapes the general poverty of the subtropical gyres.

High thermal content

Physical factors (thermal balance, atmospheric and ocean circulation) are responsible for the accumulation, in the SPC Area to the West of 180°, of a large quantity of warm water in a thick surface layer. We have already noted that this provides the largest reservoir of heat on the planet.

A large quantity of heat is transferred continuously to the atmosphere in this region, sustaining an ascending vertical circulation fed, to a large extent, by the Pacific trade winds (Walker circulation). Here we have an excellent illustration of ocean-atmosphere interaction. The ocean surface thermal field sustains the pressure field in the lower atmospheric layers, while the pressure field controls the wind field. The trade winds determine partially the ocean thermal field in the intertropical zone (cooling by equatorial upwelling in the eastern part, entrainment force of the wind on the surface holding the accumulated warm water in the western part). No one has yet succeeded in unravelling the intricacies of these interactions, even assuming it were possible to do so, so as to arrive at a continuous and clear line of perfectly defined cause-and-effect relationships. We shall therefore simply say that the high thermal content of the surface layer in the western part of the SPC Area is one of the constituents of the climatic system of the intertropical Pacific "in a normal year", the broad outline of which we have already described.

This high thermal content of the ocean layer explains in particular why the SPC Area is one of the most notorious regions for the frequency and violence of the disturbances known as tropical cyclones. It is the first of the essential conditions that must obtain simultaneously in order for these devastating meteorological phenomena to form or, at any rate, to develop. It is acknowledged that an ocean temperature greater than 27°C at the surface is necessary for a tropical cyclone to occur. The effect of the ocean thermal content has been confirmed by the numerous studies on the most consummate of this kind of meteoric condition, the RD (rapidly deepening) cyclones, which are identified by the very high speed at which the depression deepens, exceeding 42 mb in less than 24 hours. They led to the conclusion that the presence of a layer of ocean water of at least 30 m at over 28°C was a precondition for their formation (Holliday and Thompson, 1979). The second precondition for the formation of tropical cyclones could be the existence of a cyclonic-like convergence in the lower layers, a condition to be found in the West Tropical Pacific on the shear line between the equatorial winds from the West (monsoon) and the trade winds (Gray, 1977) in each hemisphere during the summer.

For instance, according to the Annual Typhoon Report published by the Joint Typhoon Warning Centre (JTWC) in Guam from 1956 to 1976, 385 cyclones were recorded in the North-west Pacific, an average of 18 a year. They occur mainly from June to November, and are most frequent in August/September. They form in Micronesia, in particular North of 10°N, and West of 170°E. They develop in the Philippine Sea and can reach the coasts of Asia, from Vietnam to Japan.

In the South-west Pacific, cyclones occur from December to March, less frequently, on average 4 a year (Ramage, 1970), mainly South of 10°S. Most of them develop in the Coral Sea. The eastern edge of the cyclone formation can be said, in a "normal year", to be 160°W, the Cook Islands longitude, and Polynesia's exposure is then limited to its most southerly islands which lie in the paths of these disturbances where they join the turbulent current of the middle latitudes to the South-east.

In years when the ENSO phenomenon occurs, the maximum thermal content zone moves towards the Central Pacific, East of 180°. At the same time, there is an eastward movement of the cyclone formation zone. Thus, in the South Pacific, between 1956 and 1978, during the 17 warm seasons regarded as "normal" (presence of equatorial upwelling West of 180°) only 16 per cent of the 176 tropical cyclones and depressions recorded originated East of 180°, compared with 31 per cent of the 51 during the five "abnormal" seasons (57-58; 65-66; 72-73; 76-77; 77-78) (Donguy et al., 1979). It should be noted then, that cyclone frequency does not seem to be higher in the El Niño years (Ramage and Hori, 1981).

This extension of the cyclone-risk zone to the South Central Pacific in an "abnormal year" was unfortunately exemplified during the 1982-83 warm season (Toan - No. 16, February 1983). Five cyclones which originated near 10°S, East of the Marquesas (140°W), devastated French Polynesia, where it is necessary to go back to the period 1903-1906 to find comparable disasters. It is a fact that the positive ocean thermal anomalies of this exceptional ENSO attained record values (5°C in the equatorial zone in the vicinity of 140°W, 10°C on the coasts of Peru) disrupting entirely the pressure and wind fields in the Intertropical Pacific and causing throughout the tropical belt climatic imbalances that subsequently extended even to the temperate latitudes.

It is precisely for the purpose of monitoring the ocean thermal field that the Tropical Ocean and Global Atmosphere (TOGA) programme was launched. For this monitoring, only satellites are capable of taking measurements "in real time" over the vast expanses of the ocean (for instance, the weekly world charts of GOSSTCOMP (Global Operational Sea Surface Temperature Computation) of NESS of NOAA) but these measurements are confined for the moment to the surface and are handicapped by atmospheric absorption constraints (Barnett et al., 1979). Such measurements from space are quality enhanced by those taken by vessels, which provide the "sea truth" for calibration and resetting and are also capable of recording, while under way, the thermal profile at a given point down to significant depths (500 m) by means of expendable bathythermographs (XBT).

Under an ORSTOM/CNEXO/SIO agreement since 1980 a dozen ships of the network organised by the Noumea ORSTOM Centre have been fitted out. The Centre has been taking measurements and samples since 1969 on the transocean lines to the South-west Pacific (Australia and New Zealand), Asia (Singapore and Japan) and America (California and Panama) via the Central Pacific (Tahiti and Hawaii). As a result, the Pacific Tropical Monitoring (SURTROPAC) and Tropical Monitoring in Polynesia (SURTROPOL) programmes of the ORSTOM network in the Pacific have become an important part of the TOGA programme.

REFERENCES

Austin, T.S. (1960). Oceanography of the East Central Equatorial Pacific as observed during expedition Eastropic - Fishery Bulletin 168 - Vol. 60, pp 257/82.

Barnett, T.P., W.C. Patzert, S.C. Webb, B.R. Bean (1979). Climatological Usefulness of Satellite Determined Seas Surface Temperatures in the Tropical Pacific. Bulletin of the American Meteorological Society, Vol. 60, No. 3, pp 197-205.

Bjerknes, J. (1966). Survey of EL NINO 1957-58 in its Relation to Tropical Pacific Meteorology, Inter-american Tropical Tuna Commission, 12 (2), 1, 1966.

Cromwell, T. (1953). Circulation in a Meridional Plane in the Central Equatorial Pacific. J. Mar. Res. 12, pp 196-213.

Denham, R.N., R.W. Bannister, K.M. Guthrie, D.G. Browning, F.G. Crook (1981). Some Hydrological Features of the South Fiji Basin. New Zealand Journal of Marine and Freshwater Research, Vol. 15, pp 299-306.

Donguy J.R., S. Begaud, R. Ebstein, B. Calvez (1979). La température de la mer et la formation des dépressions tropicales dans le Pacifique Sud. Met-mar no 103.

Donguy J.R., C. Henin, F. Rougerie (1976). Les contre-courants dans le Pacifique Tropical Sud-Ouest. Cah. ORSTOM. Ser., Oceanogr., Vol. 14, no 1, pp 15-26.

Donguy, J.R., C. Henin (1976). Anomalous Navifacial Salinities in the Tropical Pacific Ocean. Extract "Journal of Marine Research", Vol. 34 - 3 -, pp 335-64.

Donguy, J.R., C. Henin (1976). Relations entre les précipitations et la salinité de surface dans l'Océan Pacifique Tropical Sud-Ouest basées sur un échantillonnage de surface de 1956 à 1973. Annales Hydrographiques 5è série. Vol. 4, no. 2, pp 53-9.

Donguy, J.R., C. Henin (1978). Surface Salinity Fluctuations between 1956 and 1973 in the Western South Pacific Ocean. Journal of Physical Oceanography, Vol. 8, no. 6, pp 1132-4.

Donguy, J.R., C. Henin, F. Jarrige, F. Rougerie (1974). Esquisse dynamique et hydrologique du Pacifique Central Sud. Cah. ORSTOM. Sér., Oceanogr., Vol. XII, no 2, pp 129-139.

Donguy, J.R., C. Henin, A. Morliere, J.P. Rebert (1982). Thermal Changes in the Western Tropical Pacific in Relation to the Wind Field. Deep-Sea Research, Vol. 29, no. 7A, pp 869-882.

Dorman, C.E., R.H. Bourke (1979). Precipitation over the Pacific Ocean 30°S to 60°N. Month. Weather Rev., Vol. 107, July, pp 897-910.

Firing, E., C. Fernander, J. Miller (1981). Profiling Current Meter Measurements from the NORPAX Hawaii to Tahiti shuttle experiment. Hawaii Institute of Geophysics HIG. 81-2.

Gray, W.M. (1977). Tropical Cyclone Genesis in the Western North Pacific. J. Meteor. Soc., Japan (ser. II), 55, pp 465-482.

Hamon, B.U. (1970). Western Boundary Currents in the South Pacific. Scient. Explor. of the South Pacific, pp 50-9.

Hellerman, S. (1967). An updated Estimate of the Wind Stress on the World Ocean. Monthly Weather Review, Vol. 95, no 9, pp 607-614.

Henin, C., J.R. Donguy (1979). Sea Surface Salinity and Temperature Anomalies between New-Caledonia and Japan (1969-78), pp 321-31 (ORSTOM/NOUMEA).

Henin, C., J.R. Donguy (1980). Heat Content Changes within the mixed Layer of the Equatorial Pacific Ocean. Extract "Journal of Marine Research", Vol. 38-4, pp 767-79 (ORSTOM/NOUMEA).

Hidaka, K. (1958). Computation of the Wind Stresses over the Oceans. Rec. Oceanogr. Works Japan, Vol. 4, no. 2, pp 77-123.

Hisard, P., J. Merle, B. Voituriez (1970). The Equatorial Under-Current at 170°E in March and April, 1967, J.Mar. Res., 28, pp 281-303.

Holliday, C.R., A.H. Thompson (1979). Climatological Characteristics of rapidly intensifying Typhoons. Month. Weather Rev., Vol. 107, no 8, pp 1022-1034.

Jarrige, F. (1968). On the Eastward Flow of Water in the Western Pacific South of the equator, J.Mar. Res., 26, pp 286-289.

Kenyon, K.E. (1983). Sections along 35°N in the Pacific. Deep-Sea Research, Vol. 30, no 4A, pp 349-69.

Knauss, J.A. (1961). The Structure of the Pacific Equatorial Counter-Current. J. Geophys. Res., 67, pp 3943-3954.

Knauss, J.A. (1962). On some Aspects of the deep Circulation of the Pacific. J. Geophys. Res. 67 (10), pp 3943-3954.

Knauss, J.A. (1963). Equatorial Current Systems. The Sea, Vol. 2, M.N. Hill, Ed., John Wiley and Sons, London.

Knox, R.A., Halpern (1981). Variability of Winds and Upper Ocean Currents near 0°, 152°W during the Shuttle Experiment, E.O.S. 62 P 890.

Kutsuwada, K., K. Sakurai (1982). Climatological Maps of Wind Stress Field over the North Pacific Ocean. The Oceanographical Magazine, Vol. 32, no. 1-2, pp 25-48.

Lisitzin, A.P. (1970). Sedimentation and Geochemical Considerations. In: "Scientific Exploration of the South Pacific". National Academy of Sciences, Washington, D.C., pp 89-132.

Magnier, Y., H. Rotschi, P. Rual, C. Colin (1973). Equatorial Circulation in the Western Pacific (170°E). Progress in Oceanography, B.A. Warren ed., Vol. 6, 1973 pp 29-46.

Mantyla, A.W., J.L. Reid (1978). Measurements of Water Characteristics at depths greater than 10 km in the Marianas Trench. Deep-Sea Research, Vol. 25, pp 169-73.

Mantyla, A.W., J.L. Reid (1983). Abyssal Characterisitics of the World Ocean Waters. Deep-Sea Research, Vol 30. no 8A, pp 805-833.

Masuzawa, J. (1964). Flux and Water Characteristics of the Pacific North Equatorial Current. Studies on Oceanography, D. Yoshida ed., University of Washington Press, pp 121-128.

Masuzawa, J., Akiyama, Kawarada, Sawara (1970). Preliminary Report of the Ryofu Maru Cruise Ry7001 in January-March 1970, The Oceanographical Magazine Vol. 22, no 1, pp 1-25, Japan Meteorological Agency.

Masuzawa, J., K. Nagasaka (1975). The 137°E Oceanographic Section. Journal of Marine Research Supplement, Vol. 33, pp 109-16.

Merle, J., H. Rotschi, B. Voituriez (1969). Zonal circulation in the Tropical Western South Pacific at 170°E. Bulletin of the Japanese Society of Fisheries Oceanography. Special Number, pp 91-98.

Meyers, G. (1982). Inter-annual Variation in Sea Level near Truk Island. A bi-modal seasonal Cycle. Journal of Physical Oceanography, Vol. 12, no 11, pp 1161-8, American Meteorological Society.

Montgomery, R.B. (1937). A suggested Method for Representing Gradient Flow on Isentropic Surfaces. Bull. Amer. Met. Soc., 18, pp 210-12.

Montgomery, R.B., E.D. Stroup (1962). Equatorial Waters and Currents at 150°W in July-August 1952. Studies, no 1, the John Hopkin's Press.

Nitani, H. (1972). Beginning of the Kuroshio - Kuroshio, its physical Aspects. University Tokyo Press, Tokyo, 517, pp 129-163.

Oudot, C., B. Wauthy (1976). Upwelling et dôme dans le Pacifique Tropical occidental: distributions physico-chimiques et biomasse végétale. Cah. ORSTOM Sér Océanogr., Vol. XIV, no 1, pp 27-48.

Oudot, C, P. Rual, B. Wauthy (1979). Upwelling équatorial du Pacifique Ouest; distributions physico-chimiques et biomasse végétale durant une poursuite de drogue dérivante, Cahiers de l'Indo-Pacifique, Vol 1, no 1, pp 39-81.

Queney, P. (1974). Eléments de météorologie. Masson et Cie. (Ed.) PARIS.

Ramage, C.S. (1970). Meteorology of the South Pacific Tropical and middle Latitudes. Scientific Exploration of the South Pacific, pp 16-29.

Ramage, C.S., A.M. Hori (1981). Meteorological aspects of EL NINO. Monthly Weather Review, Vol. 109, no 9, pp 1827-35.

Reed, R.K., W.P. Elliott (1977). A Comparison of Oceanic Precipitation as measured by Gauge and assessed from Weather Reports. J. Appl. Meteor., 16, pp 983-986.

Reid, J.L. (1959). Evidence of a South Equatorial Counter-Current in the Pacific Ocean. Nature, 184, pp 209-210.

Reid, Jr., J.L. (1962). On Circulation, Phosphate-phosphorus Content and Zooplankton Volumes in the upper Part of the Pacific Ocean. Limnology and Oceanography No 7, pp 287-306.

Reid, Jr., J.L. (1965). Intermediate Water of the Pacific Ocean. The John Hopkin's Press Baltimore.

Reid, J.L. (1969). Sea Surface Temperature, Salinity and Density of the Pacific Ocean in Summer and in Winter. Deep-Sea Res. Supp., Vol. 16, pp 215-224.

Reid, J.L. (1969). Preliminary Results of the Measurement of deep Currents in the Pacific Ocean. Nature, 221 (5183), P 843.

Reid, J.L., R.S. Arthur (1975). Interpretation of Maps Geopotential Anomaly for the deep Pacific Ocean. Journal of Marine Research Supplement, Vol. 33, pp 37-52.

Rotschi, H. (1973). Hydrology at 170°E in the South Pacific. Oceanography of the South Pacific 1972 comp. R. Fraser. New-Zealand National Commission for UNESCO, pp 113-28.

Rotschi, H., Ph. Hisard, F. Jarrige (1972). Les eaux du Pacifique Occidental à 170°E entre 20°S et 4°N. Travaux et Documents de l'ORSTOM, no. 19.

Rotschi, H., B. Wauthy (1973). Hydrologie et dynamique dans le Pacifique occidental tropical en relation avec les mécanismes de fertilisation. Bull. Mus. Nat. Hist. Nat., PARIS, 3è série, no 147, Eco. Gen. 3.

Rougerie, F. (1981). Evaporation and Salinity Survey in French Polynesia. Tropical Ocean Atmosphere Newsletter no 7, July 1981.

Rougerie, F., L. Marec, E.P. Picard (1982). Caractéristiques hydroclimatiques de la zone marine polynésienne pendant l'année 1981, ORSTOM/Polynésie Française Océanographie. Notes et documents no 18.

Rual, P. (1969). Courants équatoriaux profonds. Deep-Sea Res., Vol. 16, pp 387-390.

Sadler, J.C., L. Oda, B.J. Kilonsky (1976). Pacific Ocean Cloudiness from Satellite Observations. University of Hawaii, UHMET 76-01.

Soegiarto, A., S. Birowo (1975). Oceanography Atlas of Indonesia and its adjacent Waters. Lembaga Oceanologi Nasional, Jakarta.

Stanton, B.R. (1975). Bibliography of the Physical Oceanography of the Tasman and Coral Seas to 1974. Misc. Pub. 66 N.Z. Ocean Inst. D.S.I.R., Wellington.

Strakhov, N.M. (1960). Principles of the Theory of Lithogenesis. Izd. Akad. Nauk, Moscow.

Stroup, E.D., B.H. Kilonsky, K. Wyrtki (1981). AXBT Observations during the Hawaii/Tahiti Shuttle Experiments. University of Hawaii, Rep. UHMET 81-05, 54 pgs.

Sverdrup, H.U., M.W. Johnson, R.H. Fleming (1942). The Oceans, their Physics, Chemistry and General Biology. Prentice Hall, Inc., New York.

Tabata, S. (1975). The general Circulation of the Pacific Ocean and a brief Account of the Oceanographic Structure of the North Pacific Ocean, Part 1, Circulation and Volume Transports, Atmosphere Vol. 13, no 4, pp 133-168.

Taft, B.A., P. Kovala (1981). Vertical Section of Temperature, Salinity, Thermosteric Anomaly and Zonal Geostrophic Velocity from NORPAX Shuttle Experiment, Part I, NOAA Data Report ERL - PMEL-3, 98 Pgs.

Takahashi, T., M. Chaen, S. Ueda (1960). Report of the Kagoshima-maru IGY Cruise, 1958. "Memories of the Faculty of Fisheries Kaposime University", Vol. 8, pp 82-6.

Taylor, R.C. (1973). An Atlas of Pacific Islands Rainfall. Document of the "Office of Naval Research", H16-73-9.

Thurman, H.V. (1973). Introductory Oceanography. C.E. Merrill Publishing Company, Columbus, Ohio, USA.

T.O.A.N. - Tropical Ocean Atmosphere Newsletter. Published bimonthly since 1980 by Univ. of Washington Joint Institute for the Study of the Atmosphere and Ocean (JISAO) AK-40, Seattle, WA 98195 USA.

Tsuchiya, M. (1968). Upper Waters of the Intertropical Pacific Ocean. The John Hopkin's Press, Baltimore.

Uda, M., K. Hasunuma (1969). The Eastward Subtropical Counter-Current in the Western North Pacific Ocean. J. Oceanogr. Soc. Japan, 25, pp 201-210.

Warren, B.A. (1970). General Circulation of the South Pacific In: "Scientific Exploration of the South Pacific", pp 33-49. National Academy of Sciences, Washington D.C..

White, W.B., R. Wyle (1977). Annual and seasonal Maps of Residual Temperature in the upper Waters of the Western North Pacific from 1954-1974. Scripps Inst. Oceanogr., La Jolla, Calif. SIO Ref. 77-28.

White, W.B., G. Meyers (1982). Space/time Statistics of short-term Climatic Variability in the Western North Pacific. Journal of Geophysical Research, Vol. 87, no C3, pp 1979-1989.

Wooster, W.S. (1961). Further evidence of a Pacific South Equatorial Counter-Current. Deep-Sea Res., 8, pp 294-297.

Worthington, L.V., H. Kawaii (1972). Comparison between deep Sections across the Kuroshio and the Florida Current and Gulf Stream - Kuroshio, its physical Aspects. Univ. Tokyo Press, 517, pp 371-385.

Wyrtki, K. (1965). The annual and semi-annual Variation of Sea Surface Temperature in the North Pacific Ocean. Limnology and Oceanography, Vol. X, no 3, pp 307-13.

Wyrtki, K. (1974). Sea Level and the seasonal Fluctuations of the Equatorial Currents in the Western Pacific Ocean. Extract "Journal of Physical Oceanography", Vol. 4, no 1, pp 91-103. American Meteorological Society.

Wyrtki, K. (1975). Fluctuations of the dynamic Topography in the Pacific Ocean. J.Phys. Oceanogr., 5, pp 450-459.

Wyrtki, K. (1978). Monitoring the Strength of Equatorial Currents from XBT Sections and Sea Level. Journal of Geophysical Research, Vol. 83, no C4, pp 1935-40.

Wyrtki, K. (1981). An Estimate of Equatorial Upwelling in the Pacific. Extract "Journal of Physical Oceanography", Vol. 11, no 9, pp 1205-14. American Meteorological Society.

Wyrtki, K. (1982). The Southern Oscillation, Ocean-Atmosphere Interaction and EL NINO. Marine Technology Society Journal, Vol. 16, no 1.

Wyrtki, K., G. Eldin (1982). Equatorial Upwelling Events in the Central Pacific Journal of Physical Oceanography, Vol. 12, no 9, pp 984-8. American Meteorological Society.

Wyrtki, K., R. Kendall (1967). Transports of the Pacific Equatorial Counter-Current. J. Geophys. Res., 72, pp 2073-2076.

Wyrtki, K., B. Kilonsky (1982). Transequatorial Water Structure during the Hawaii to Tahiti Shuttle Experiment. Hawaii Institute of Geophysics - HIG - 825.

Wyrtki, K., W.G. Leslie (1980). The Mean Annual Variation of Sea Level in the Pacific Ocean. Hawaii Institute of Geophysics - HIG - 80-5.

Wyrtki, K., G. Meyers (1975). The Trade Wind Field and the Mean Annual Variation (Part 1). The Mean Field and the Mean Annual Variation. National Science Foundation and Office of Naval Research HIG 75-1, Hawaii Institute of Geophysics, Univ. of Hawaii.

Yamanaka, H., Anraku, Morita (1965). Seasonal and long-term Variations in Oceanographic Conditions in the Western North Pacific Ocean. Nankai Regional Fisheries Research Laboratory, no 22.

Yamanaka, H. (1973). Surface Currents in the Western Pacific Ocean as seen from the Drift of Tuna Longline Gear. Bulletin 9, Nov. 73. Far Seas Fish. Research Lab. Shimizu 424, Japan.

Yoshida, K., T. Kidokoro (1967). A Subtropical Counter-Current. A Prediction of Eastward Flows at lower Subtropical Latitudes. J. Oceanogr. Soc. Japan, 23, pp 231-246.

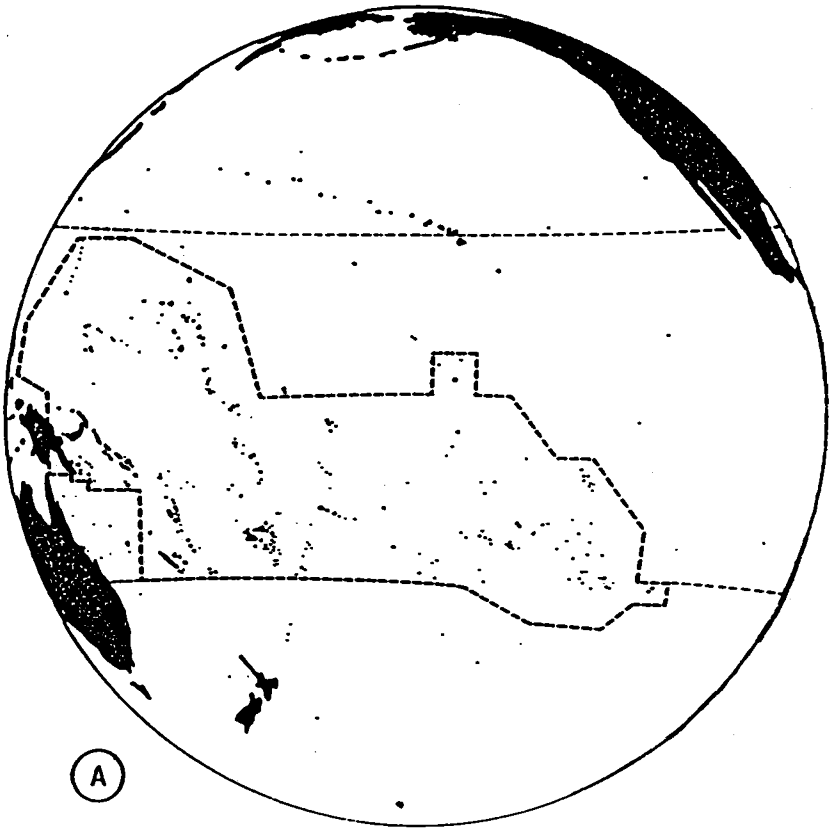
Yoshimura, H. (1972). On the Distribution of Phosphorus along 170°W Longitude in the Pacific and Antarctic Oceans. The Oceanographical Magazine, Vol. 24, no 1, pp 65-79.

FIGURES

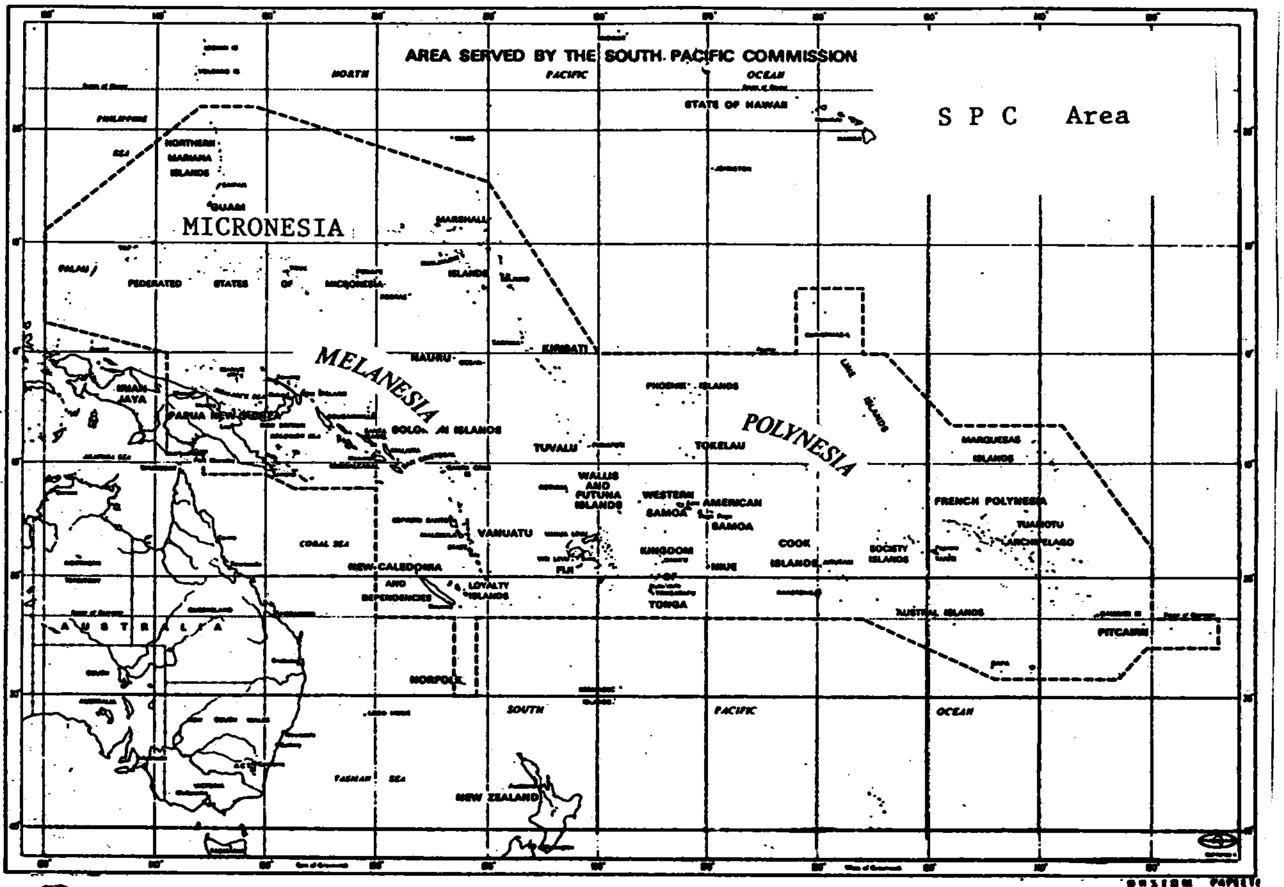
Fig. 33 Diagram of possible transports, in Sverdrups (1 Sv. = $10^6 m^3.s^{-1}$) of currents in the SPC Area.

Fig. 34 Primary production in the Pacific (gC/m^2/year).

Fig. 35 Zooplankton distribution in the Pacific (0 – 150 m; mm^3/m^3); according to Reid (1962).



(A)


AREA SERVED BY THE SOUTH PACIFIC COMMISSION
S P C Area
MICRONESIA
MELANESIA
POLYNESIA
NORTH PACIFIC OCEAN
STATE OF HAWAII
NORTHERN MARIANA ISLANDS
GUAM
MARSHALL ISLANDS
FEDERATED STATES OF MICRONESIA
NAURU
KIRIBATI
PHOENIX ISLANDS
TUVALU
TOKELAU
WALLIS AND FUTUNA ISLANDS
WESTERN SAMOA
AMERICAN SAMOA
VANUATU
FIJI
KINGDOM OF TONGA
NIUE
COOK ISLANDS
SOCIETY ISLANDS
FRENCH POLYNESIA
MARQUESAS ISLANDS
TUAMOTU ARCHIPELAGO
AUSTRAL ISLANDS
PITCAIRN
CORAL SEA
NEW CALEDONIA AND DEPENDENCIES
LOYALTY ISLANDS
NORFOLK
SOUTH PACIFIC OCEAN
TASMAN SEA
NEW ZEALAND
AUSTRALIA
PAPUA NEW GUINEA
IRIAN JAYA
SOLOMON ISLANDS
PHILIPPINE SEA
PALAU

(B)

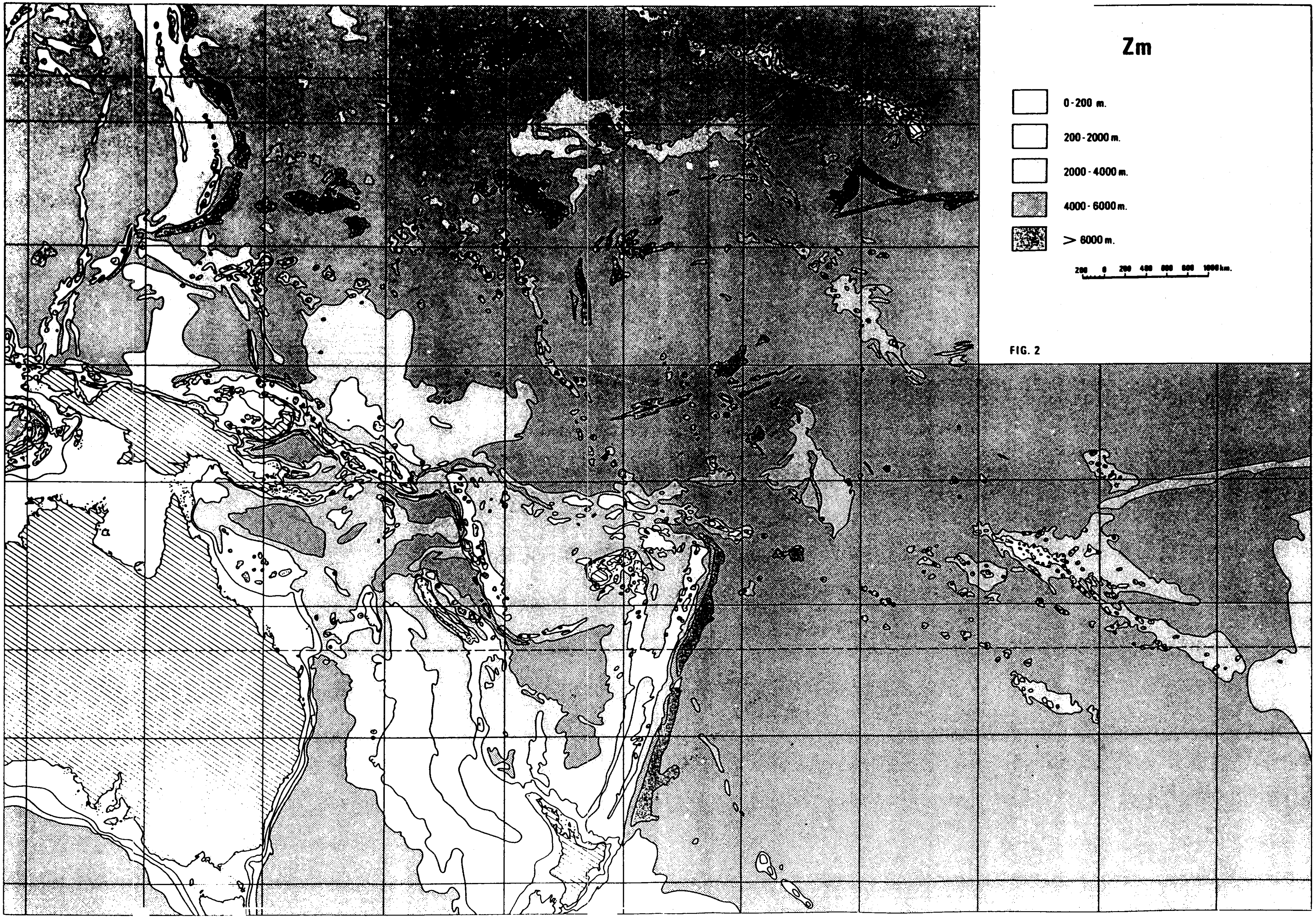

Zm
0-200 m.
200-2000 m.
2000-4000 m.
4000-6000 m.
> 6000 m.
200 0 200 400 600 800 1000 km.

FIG. 2

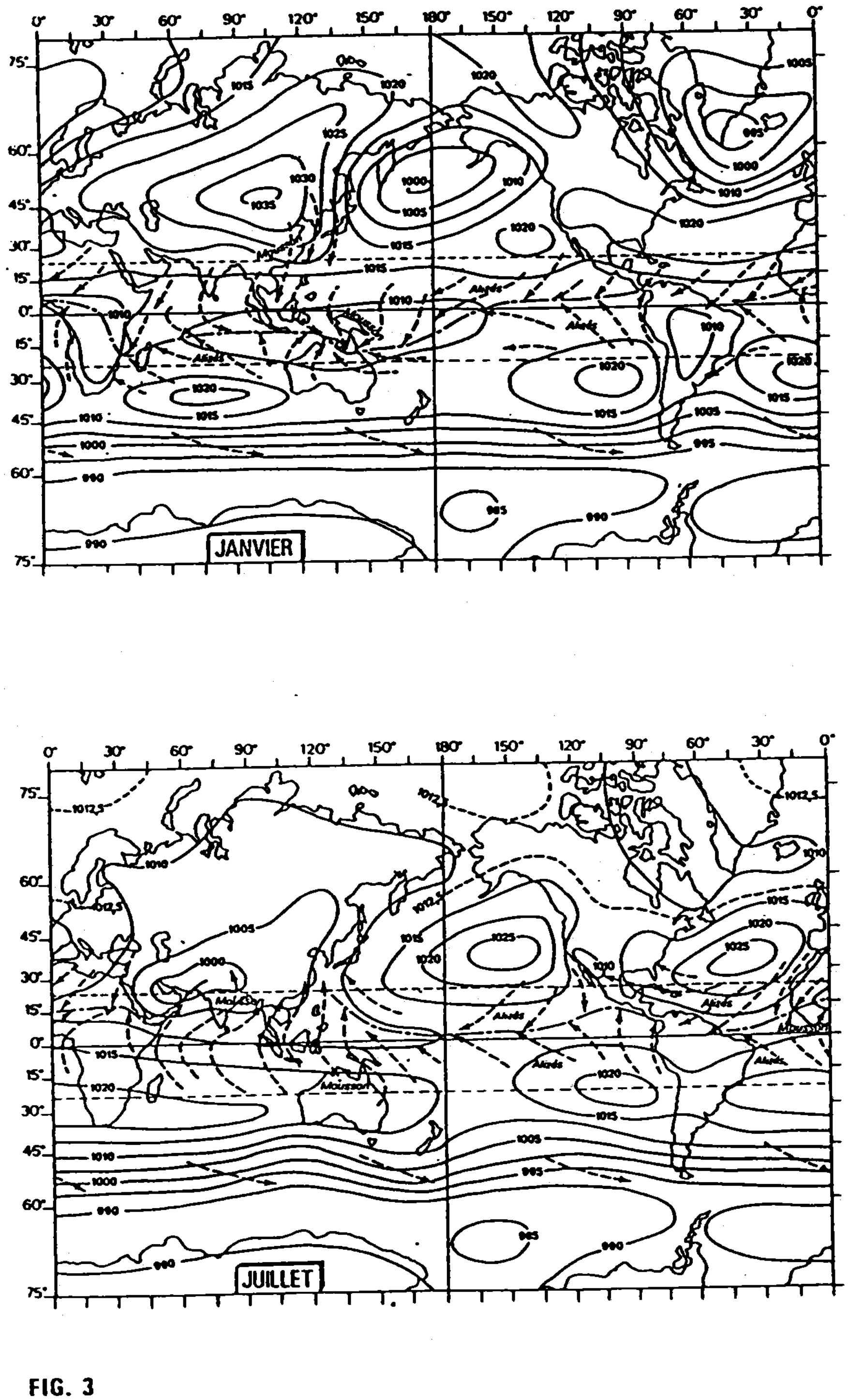
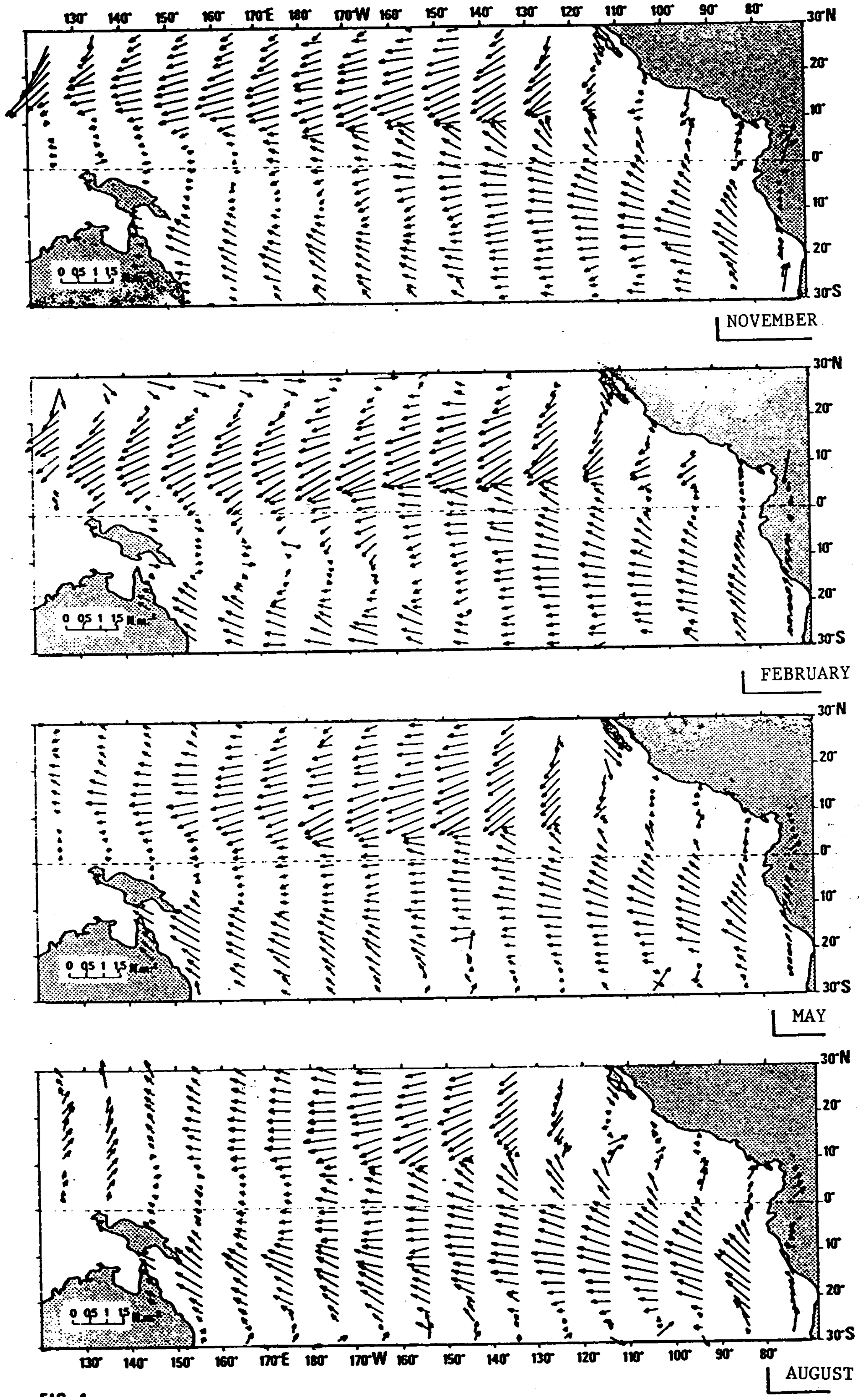
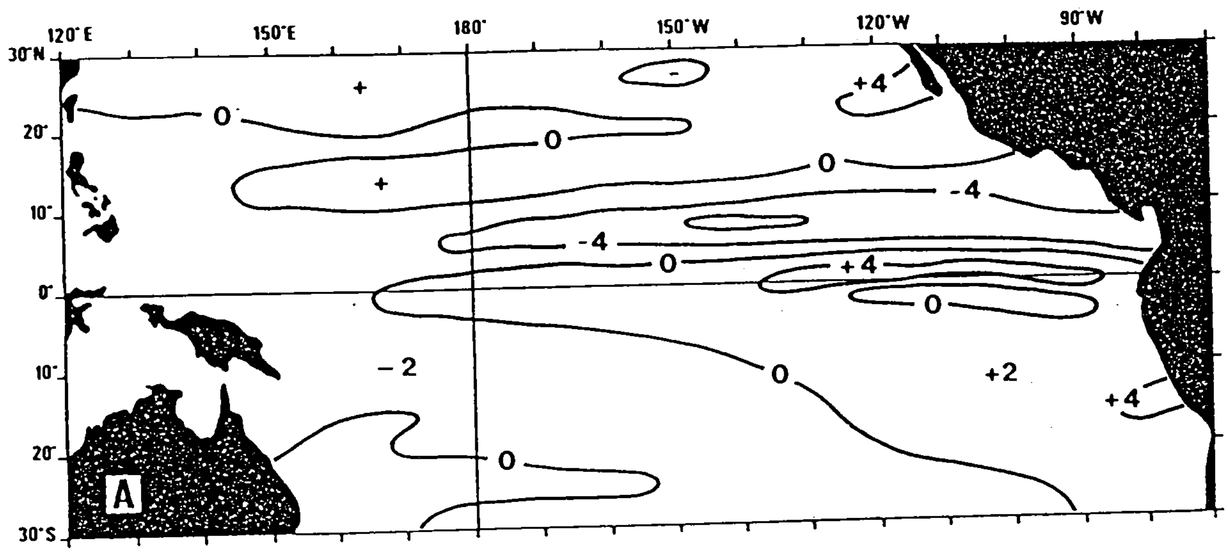


FIG. 3


130° 140° 150° 160° 170°E 180° 170°W 160° 150° 140° 130° 120° 110° 100° 90° 80°
30°N
20°
10°
0°
10°
20°
30°S
0 0.5 1 1.5 N.m⁻²
NOVEMBER
FEBRUARY
MAY
AUGUST



120° E
150°E
180°
150°W
120°W
90°W
30°N
20°
10°
0°
10°
20°
30°S
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+
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+4
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-4
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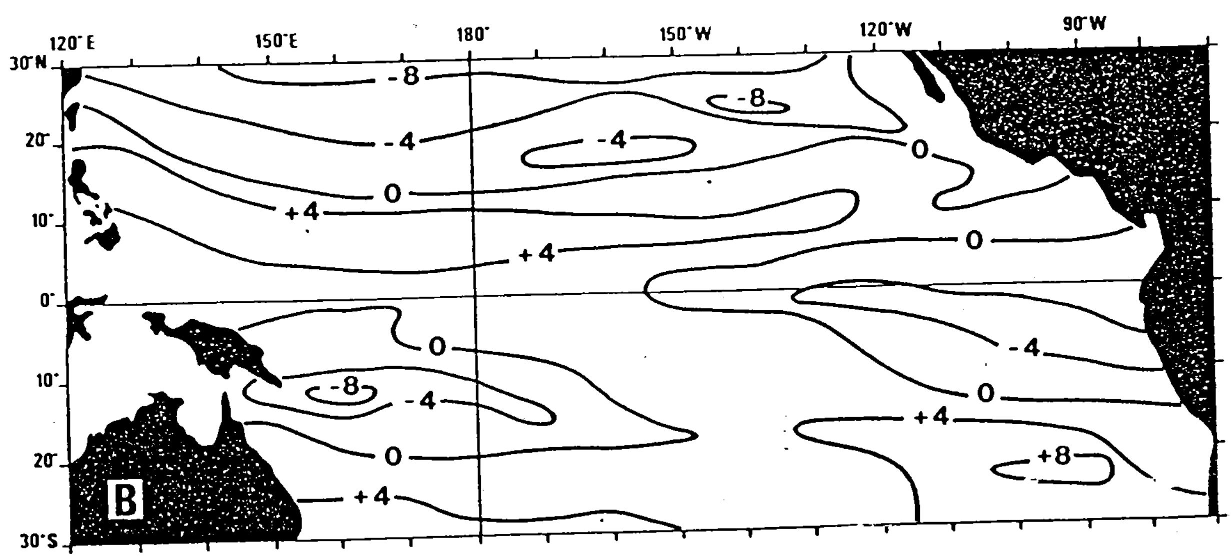
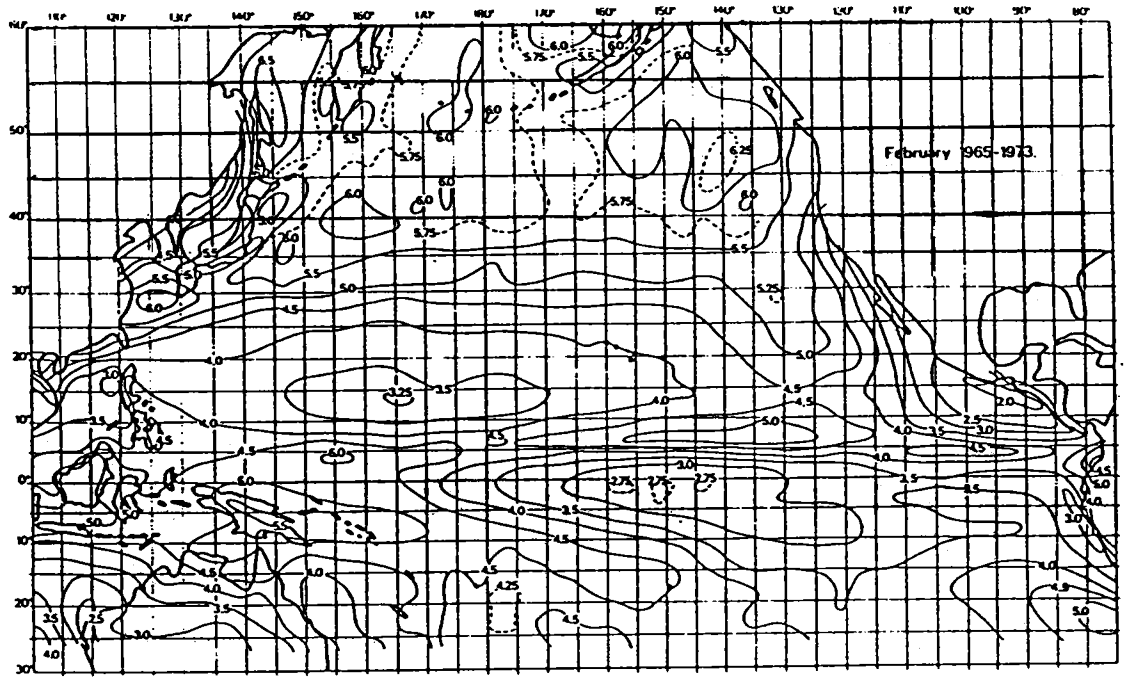
0
+2
+4
0



120° E
150°E
180°
150°W
120°W
90°W
30°N
20°
10°
0°
10°
20°
30°S
B
-8
-8
-4
-4
0
0
+4
+4
0
0
-4
-8
-4
0
+4
0
+8
+4

FIG. 5



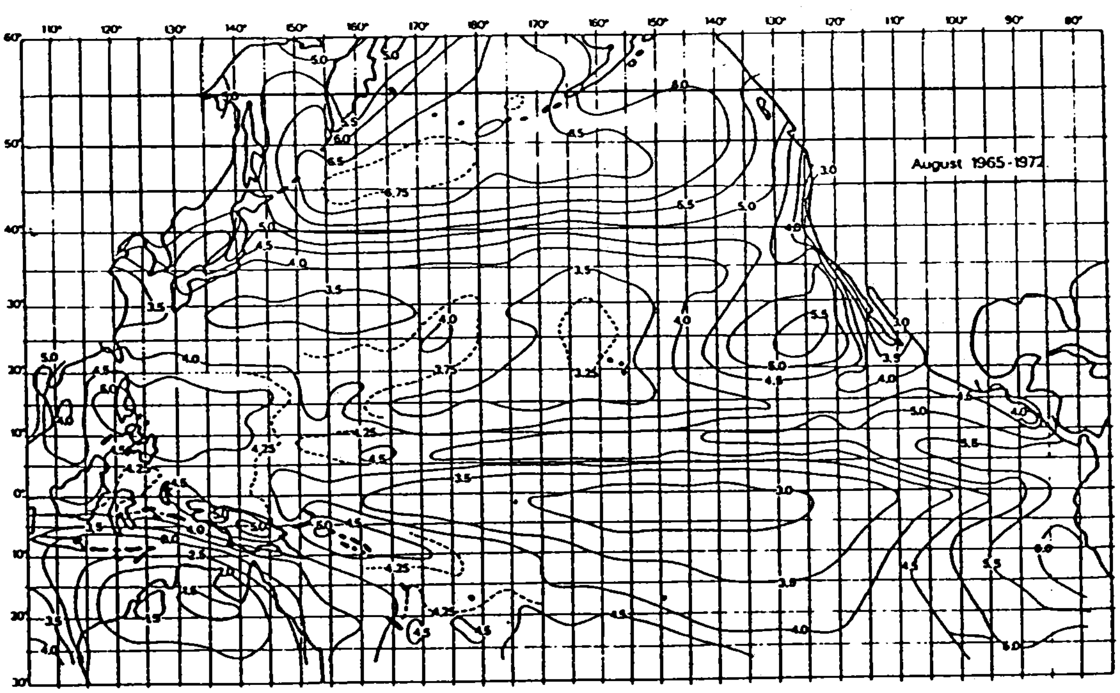
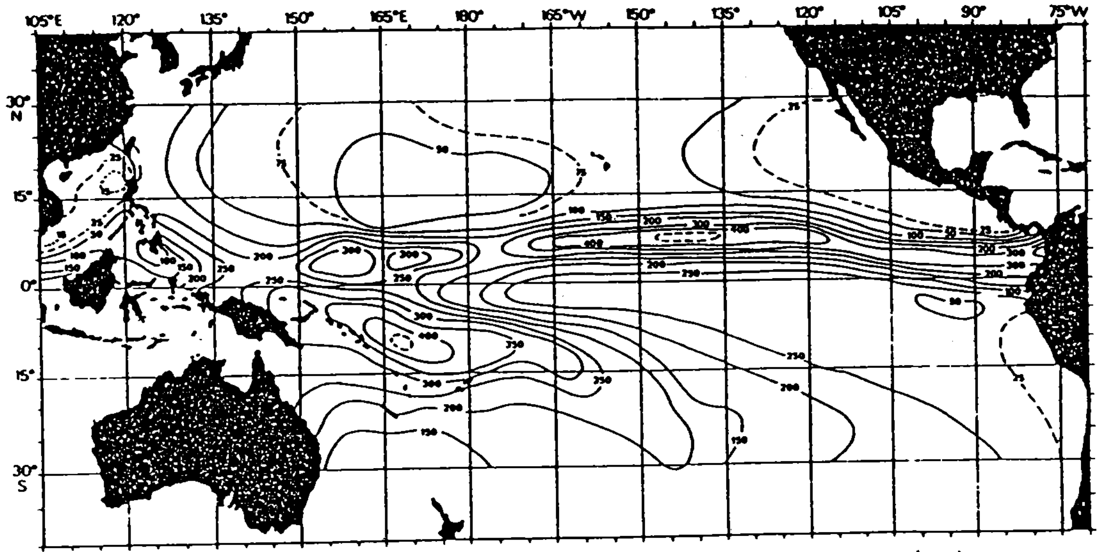
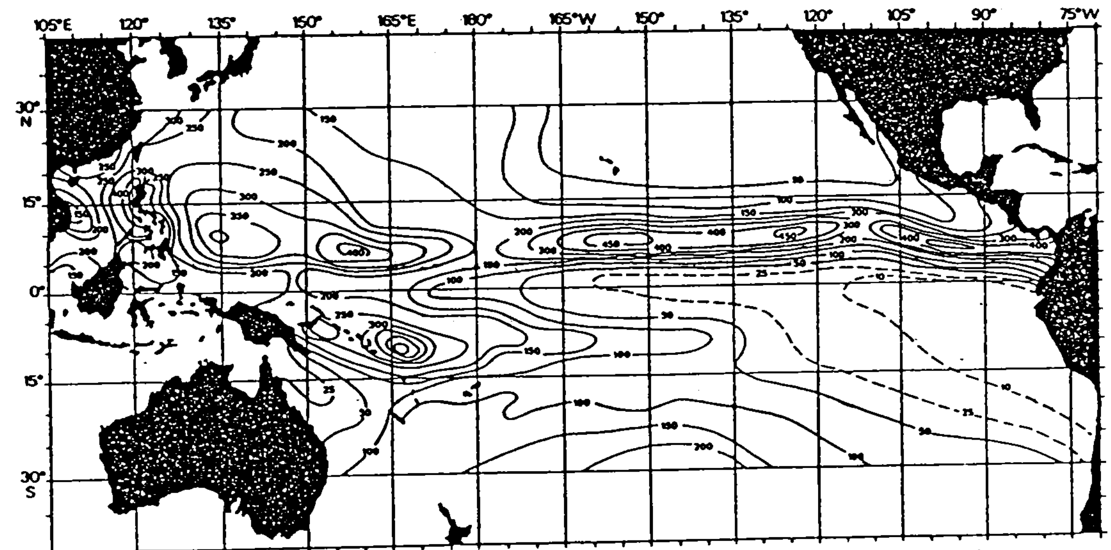


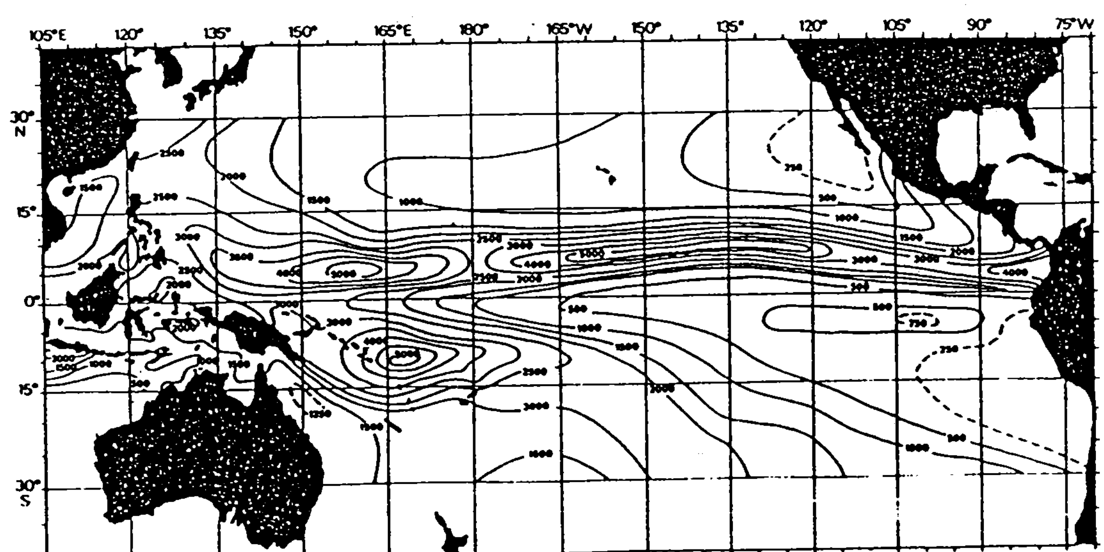
FIG. 6



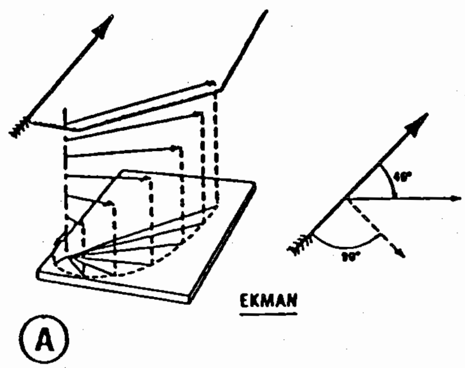
Feb (mm)

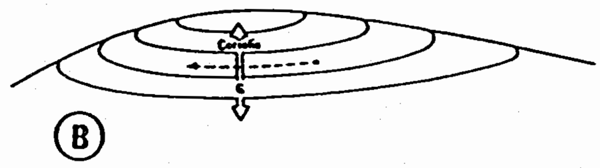


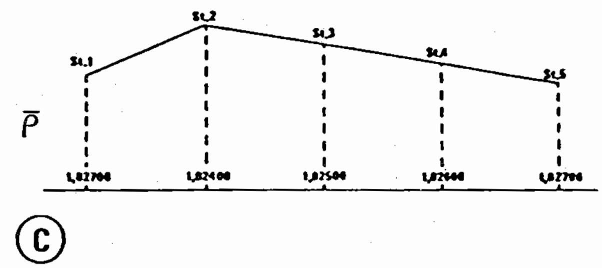
Aug (mm)

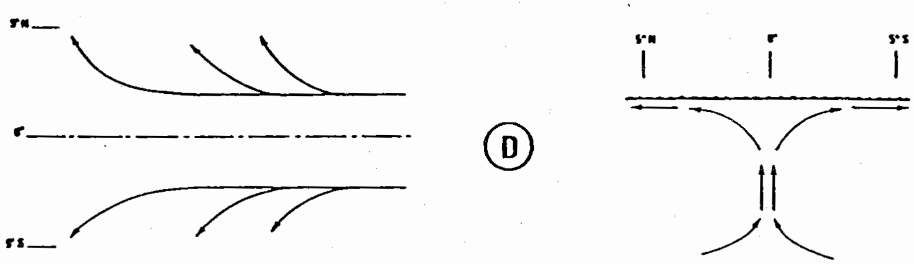


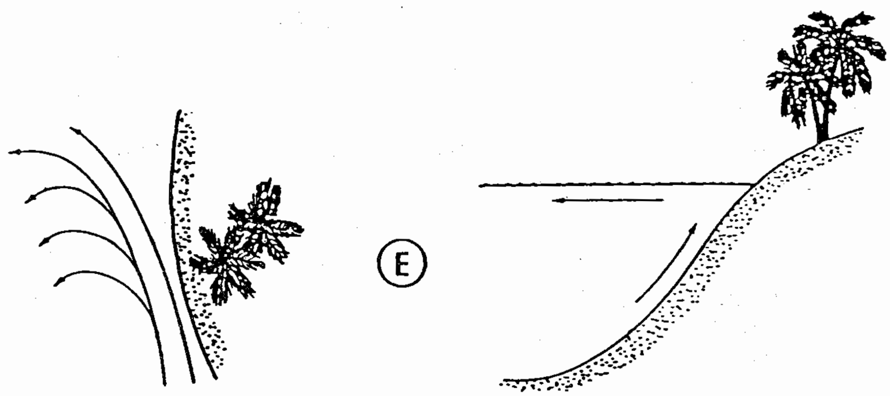
$\overline{An}$ (mm)

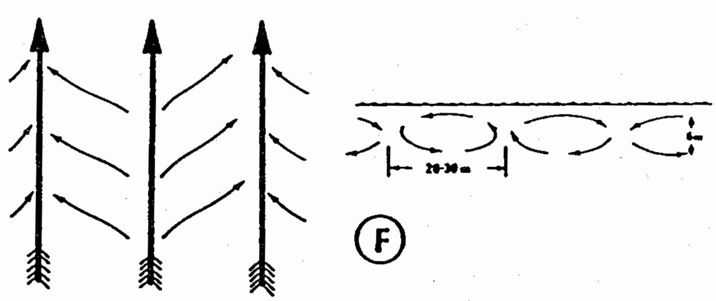

EKMAN
40°
90°
A

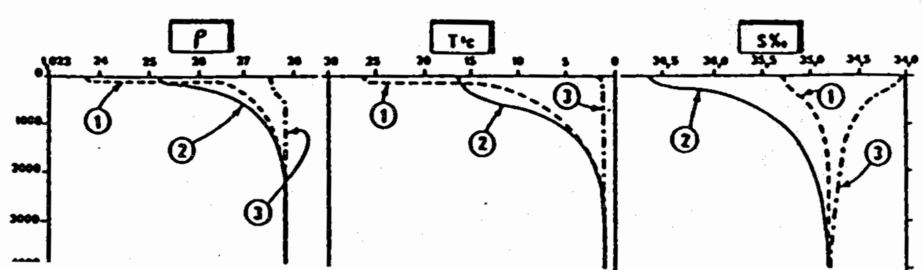

Coriolis
G
B

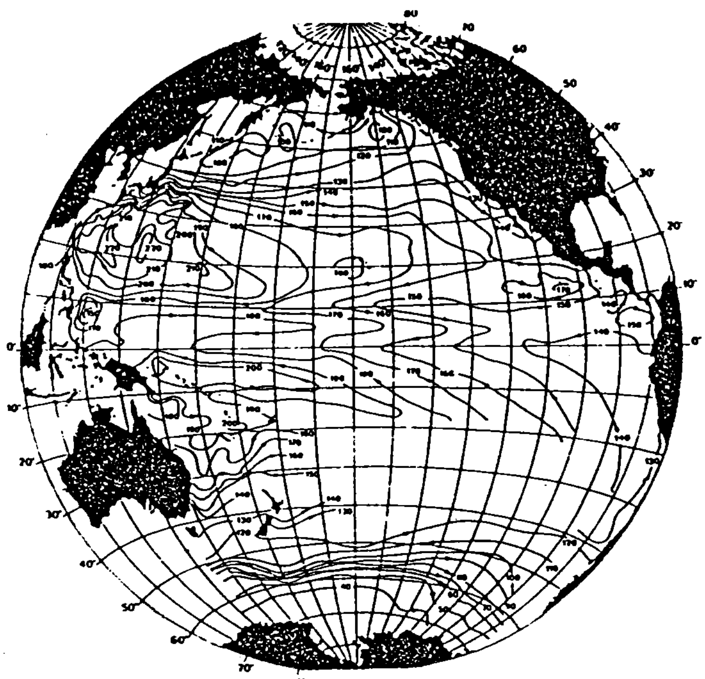

St.1
St.2
St.3
St.4
St.5
$\overline{P}$
1,02700
1,02400
1,02500
1,02600
1,02700
C


5°N
0°
5°S
5°N
0°
5°S
D

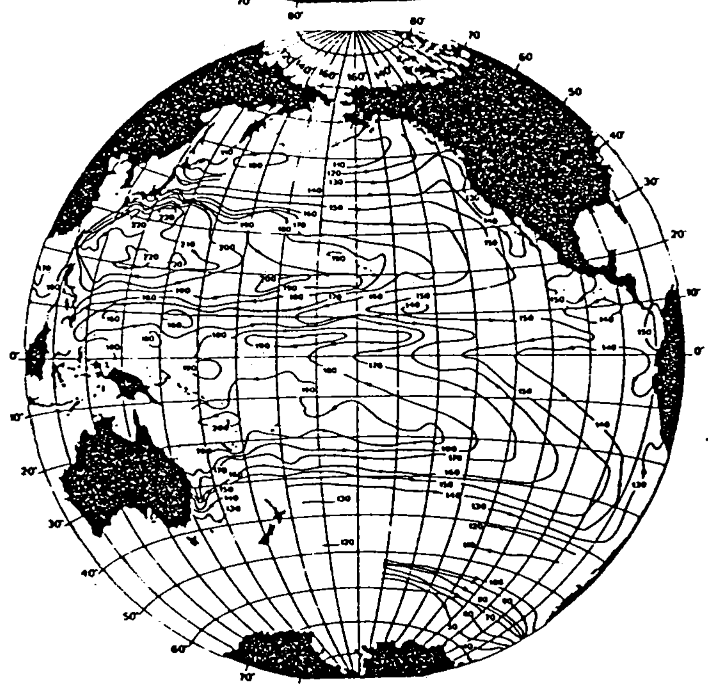

E


20-30 m
F

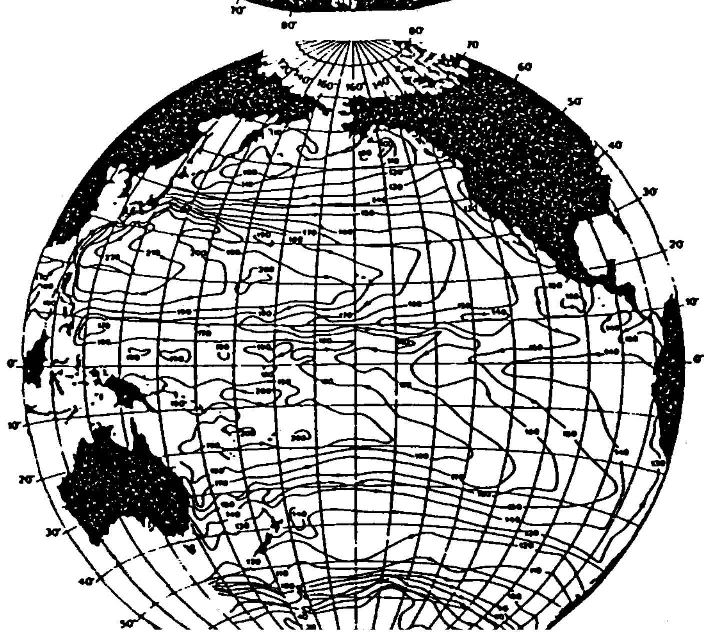

ρ
T°c
S‰
1000
2000
3000



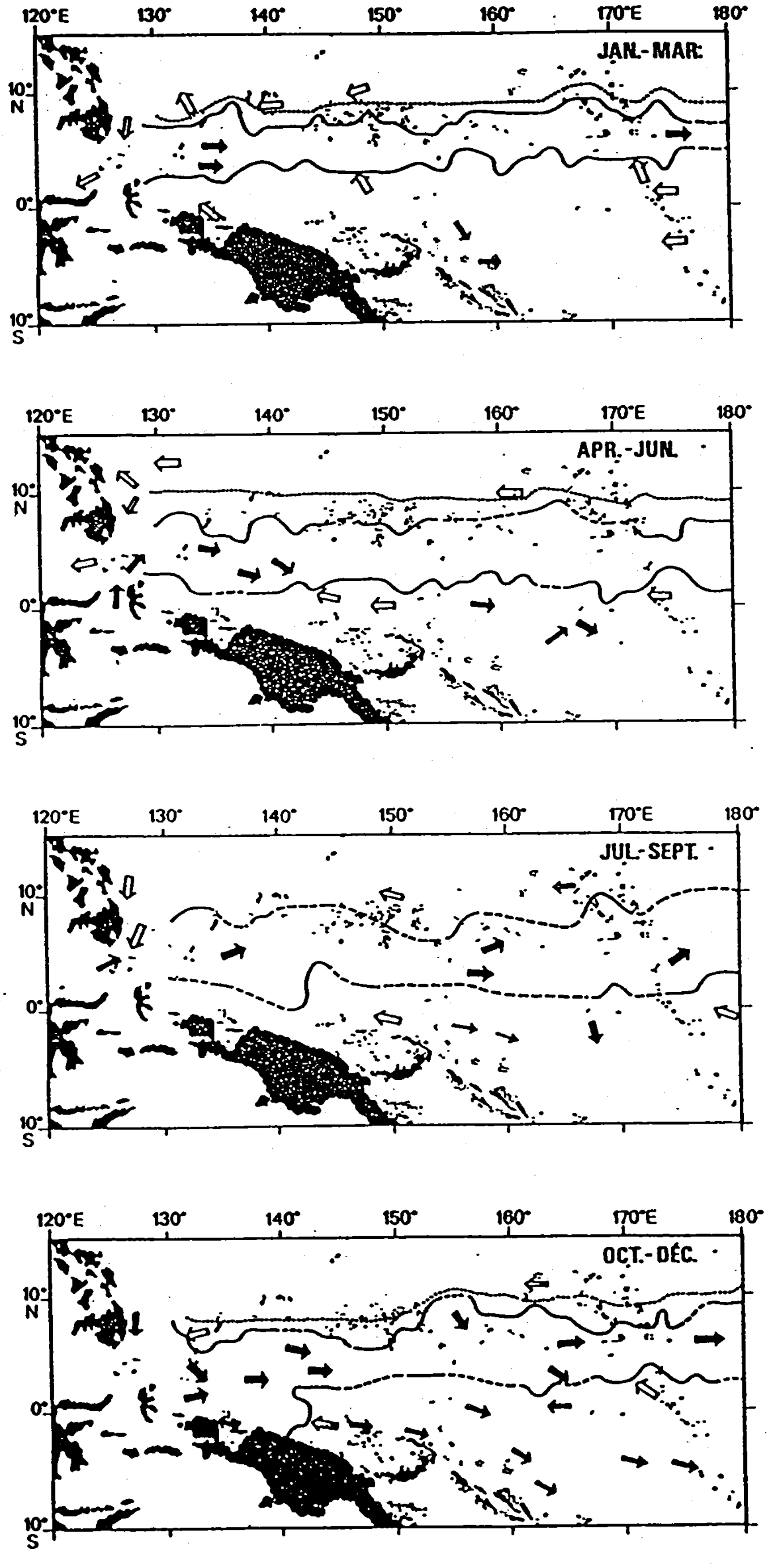
Mar.-Apr. $\overline{\text{dyn.cm.}}$ 0/1000 db.



Nov.-Dec. $\overline{\text{dyn.cm.}}$ 0/1000 db.



An $\overline{\text{dyn. cm.}}$ 0/1000db.


120°E
130°
140°
150°
160°
170°E
180°
JAN.-MAR.
10° N
0°
10° S
APR.-JUN.
JUL.-SEPT.
OCT.-DÉC.

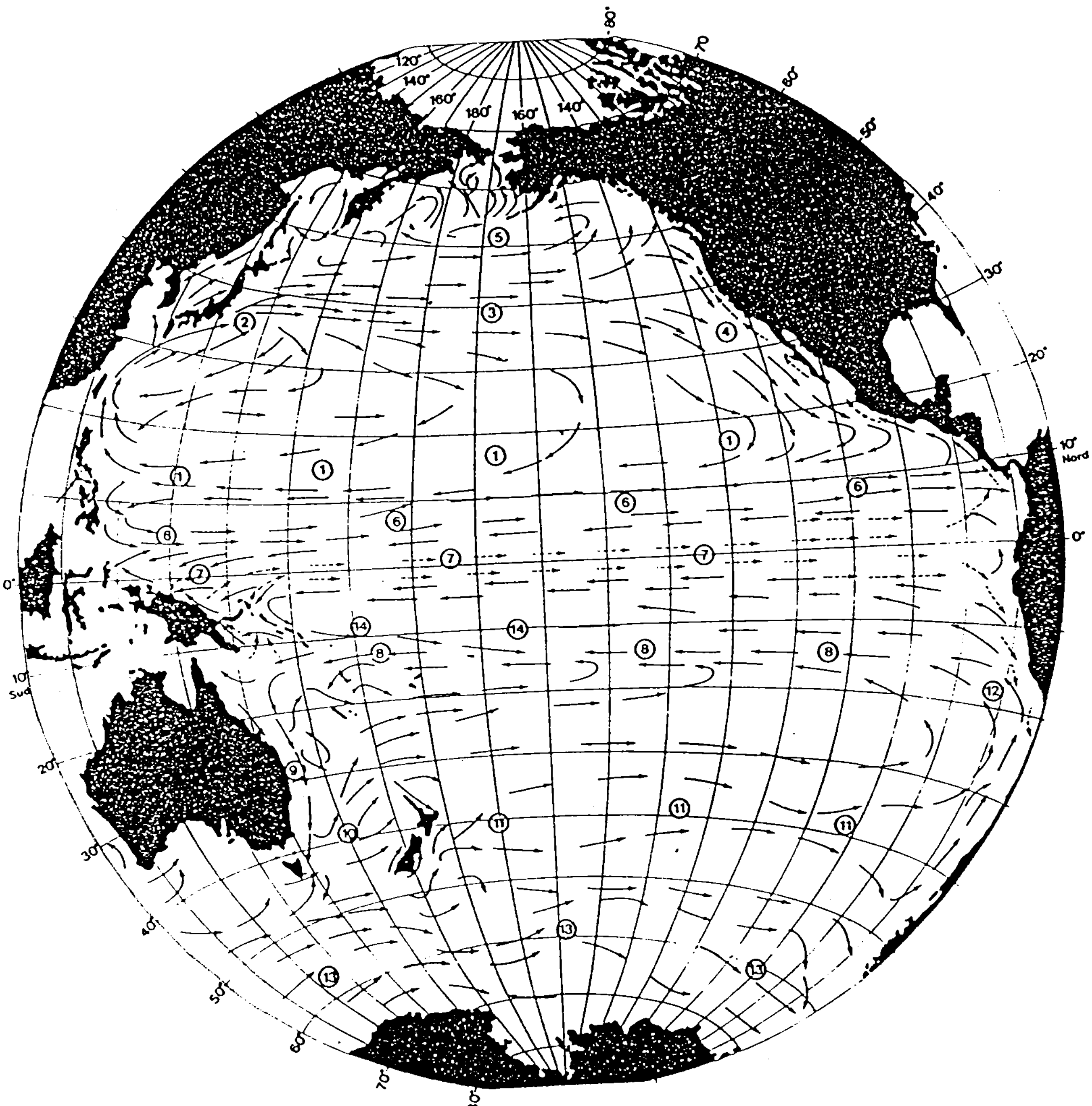
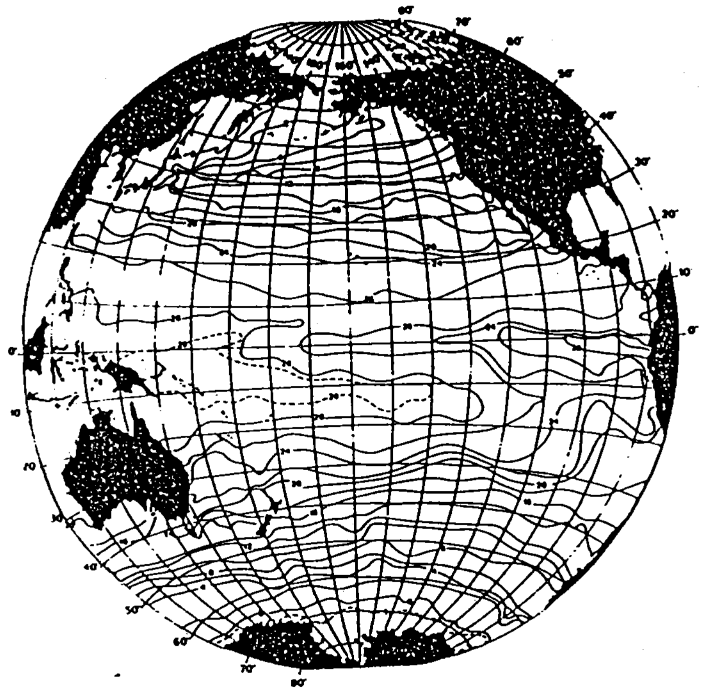
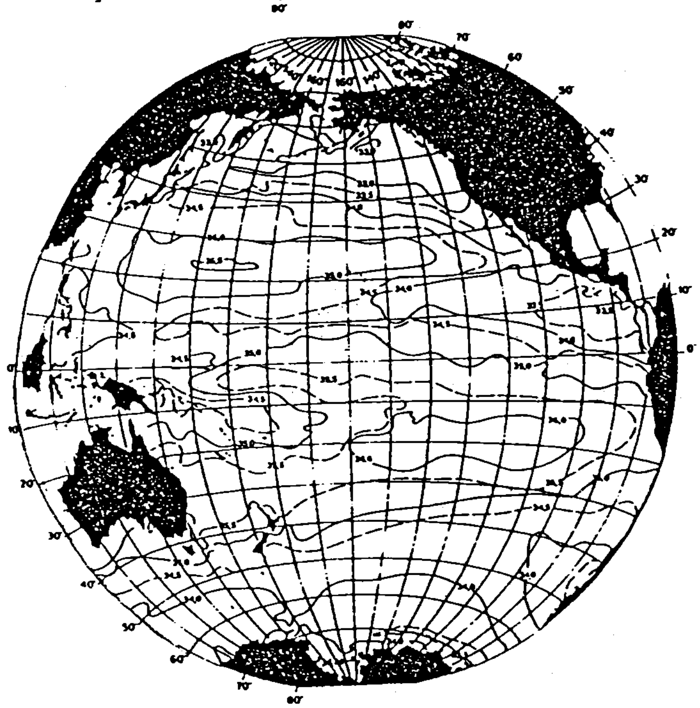

80°
70
80°
50°
40°
30°
20°
10°
Nord
0°
120°
140°
160°
180°
160°
140°
0°
10°
Sud
20°
30°
40°
50°
60°
70°
80°

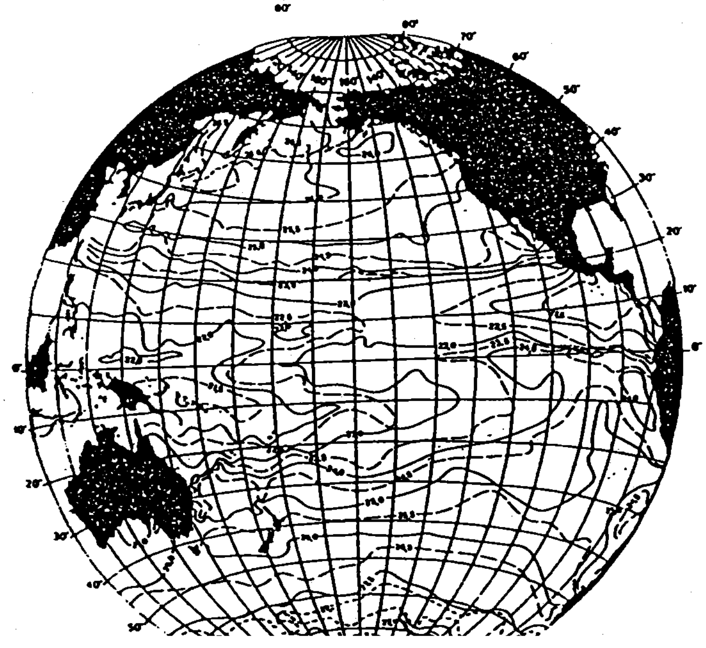
FIG. 11



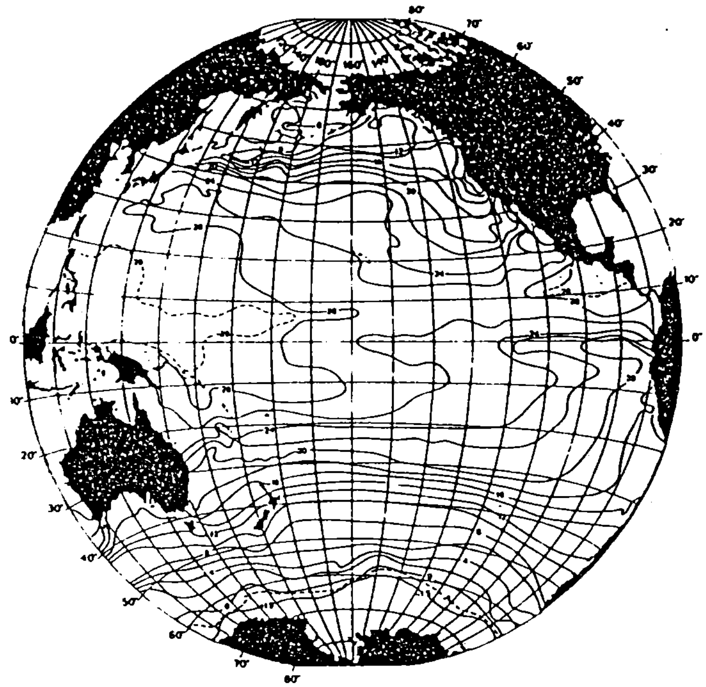
T °c
D.J.F.



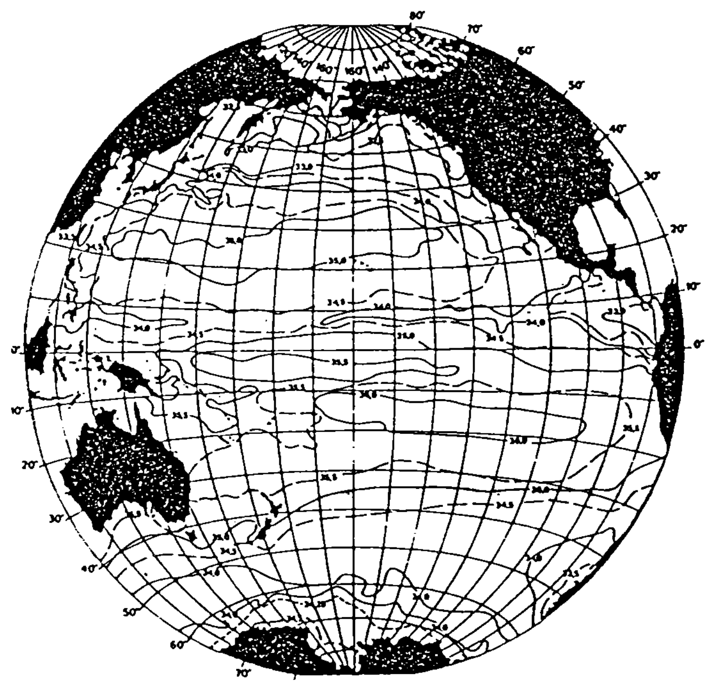
S‰
D.J.F.



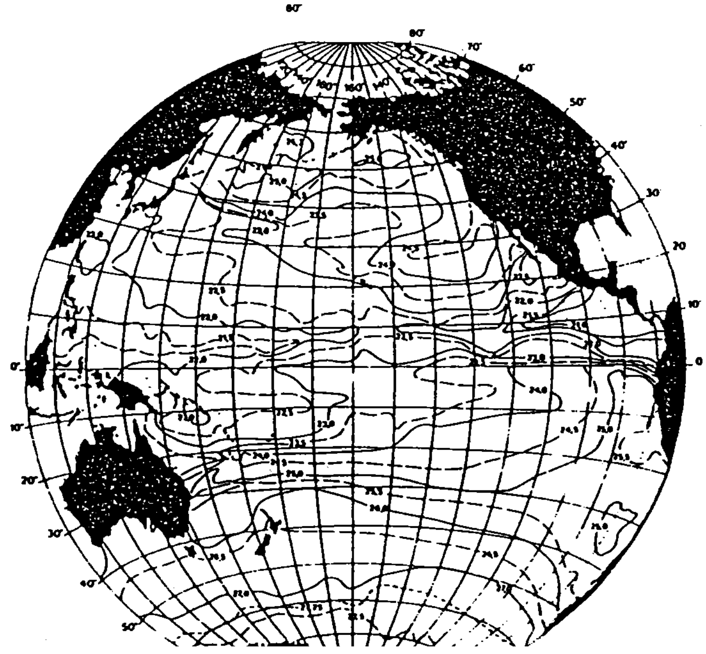
σ_t
D.J.F.



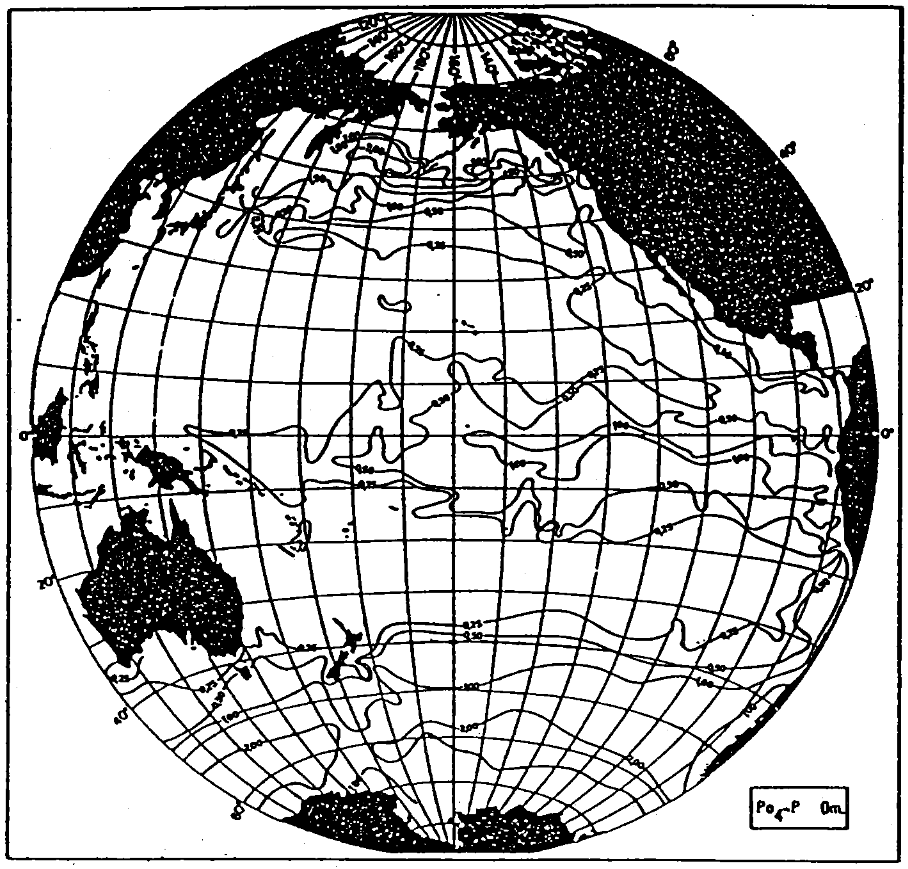
T °C
JJA



S ‰
JJA



σ_t
JJA



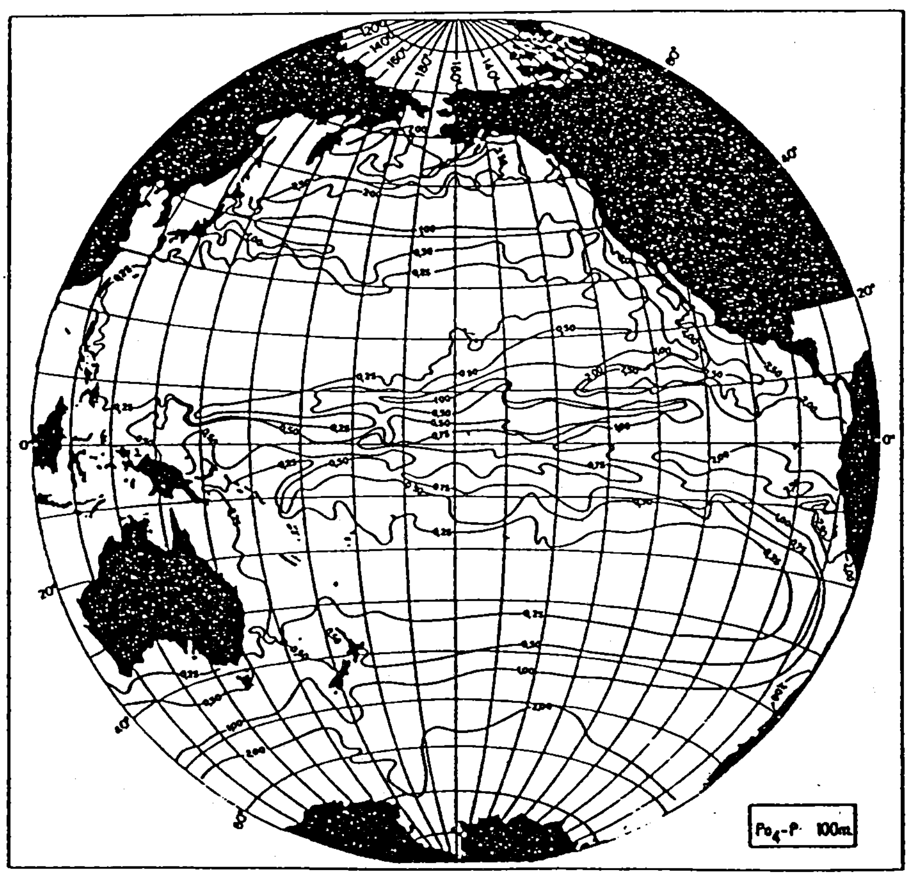
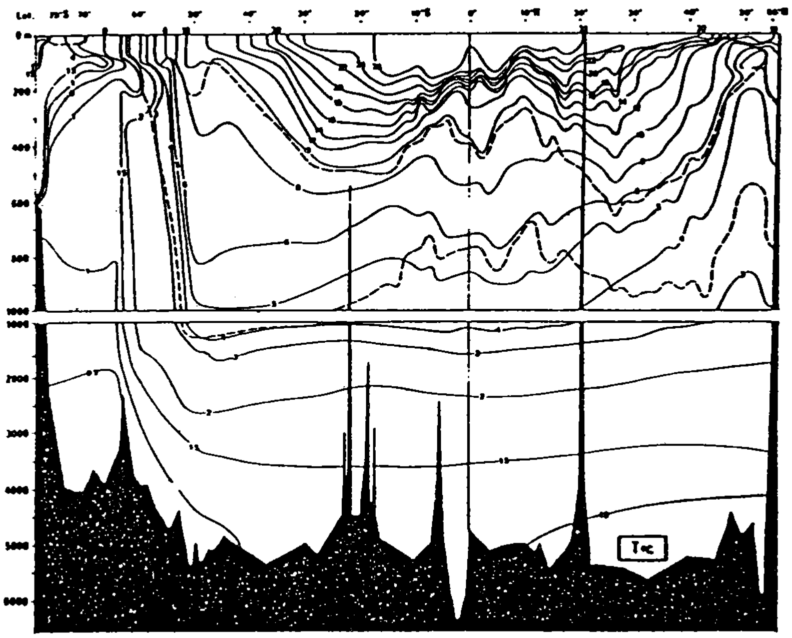
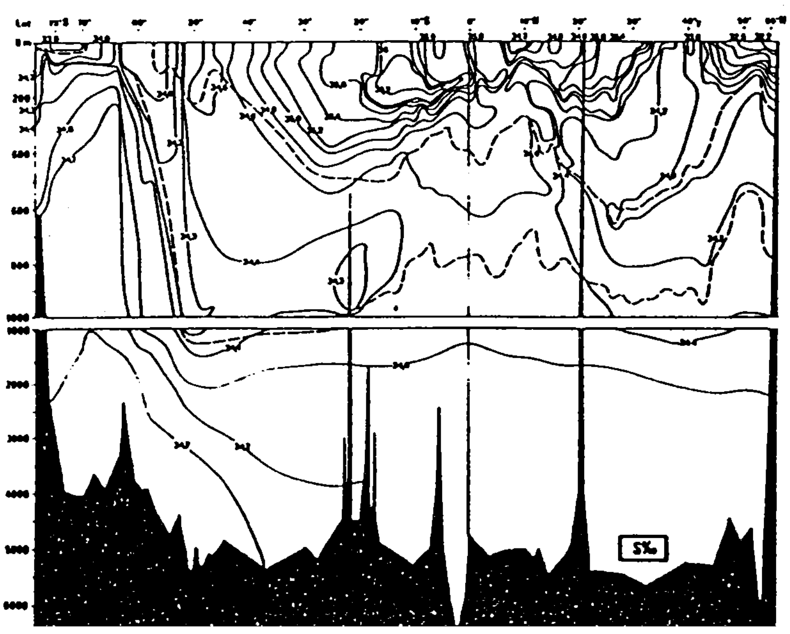
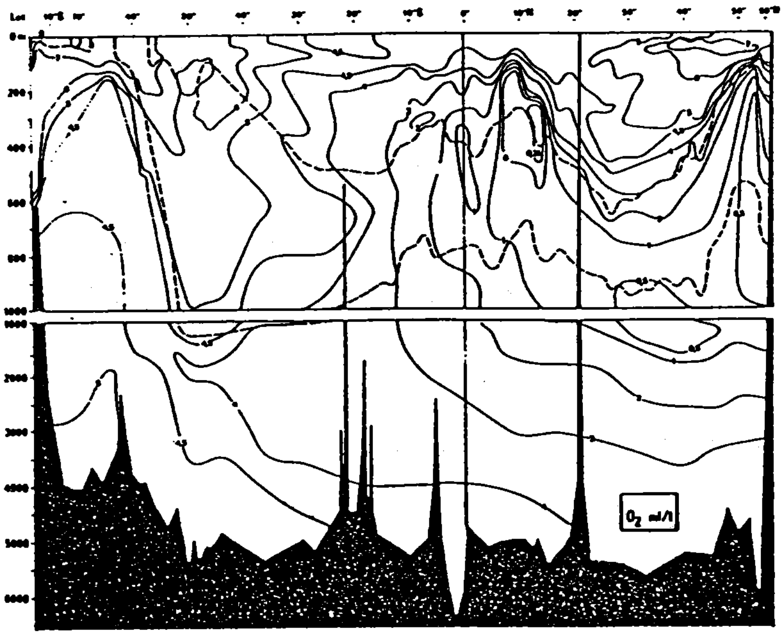


FIG. 14


d/t


T°c


S‰



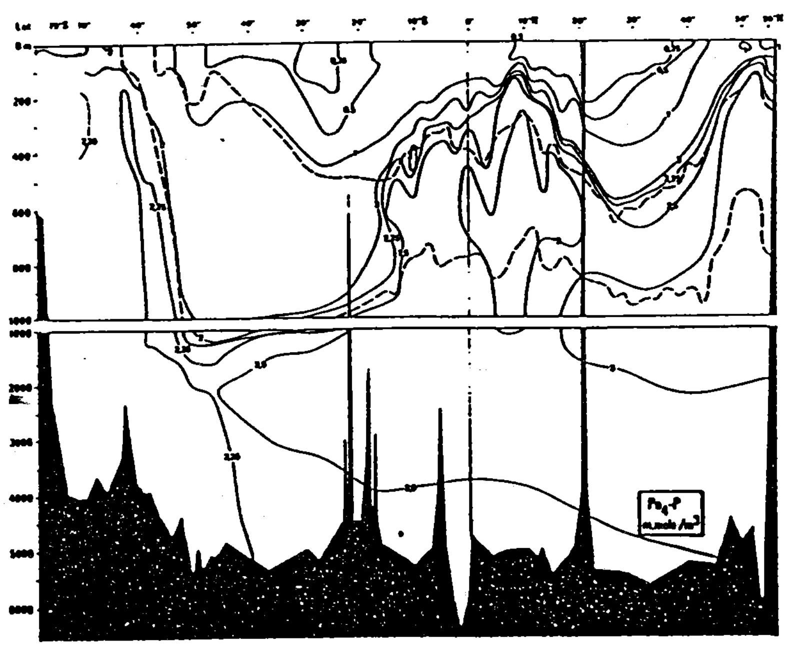


FIG. 16

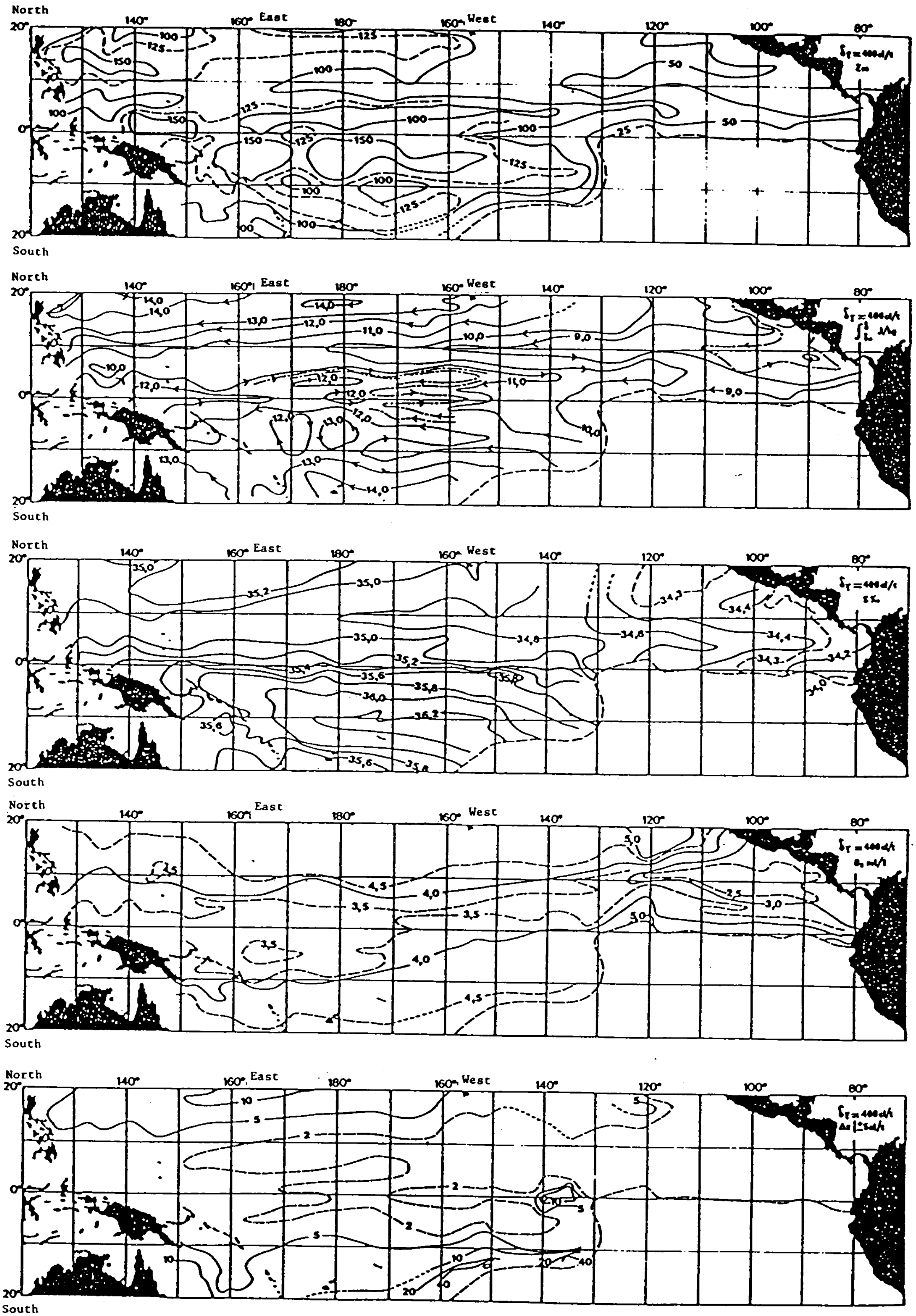


FIG. 17

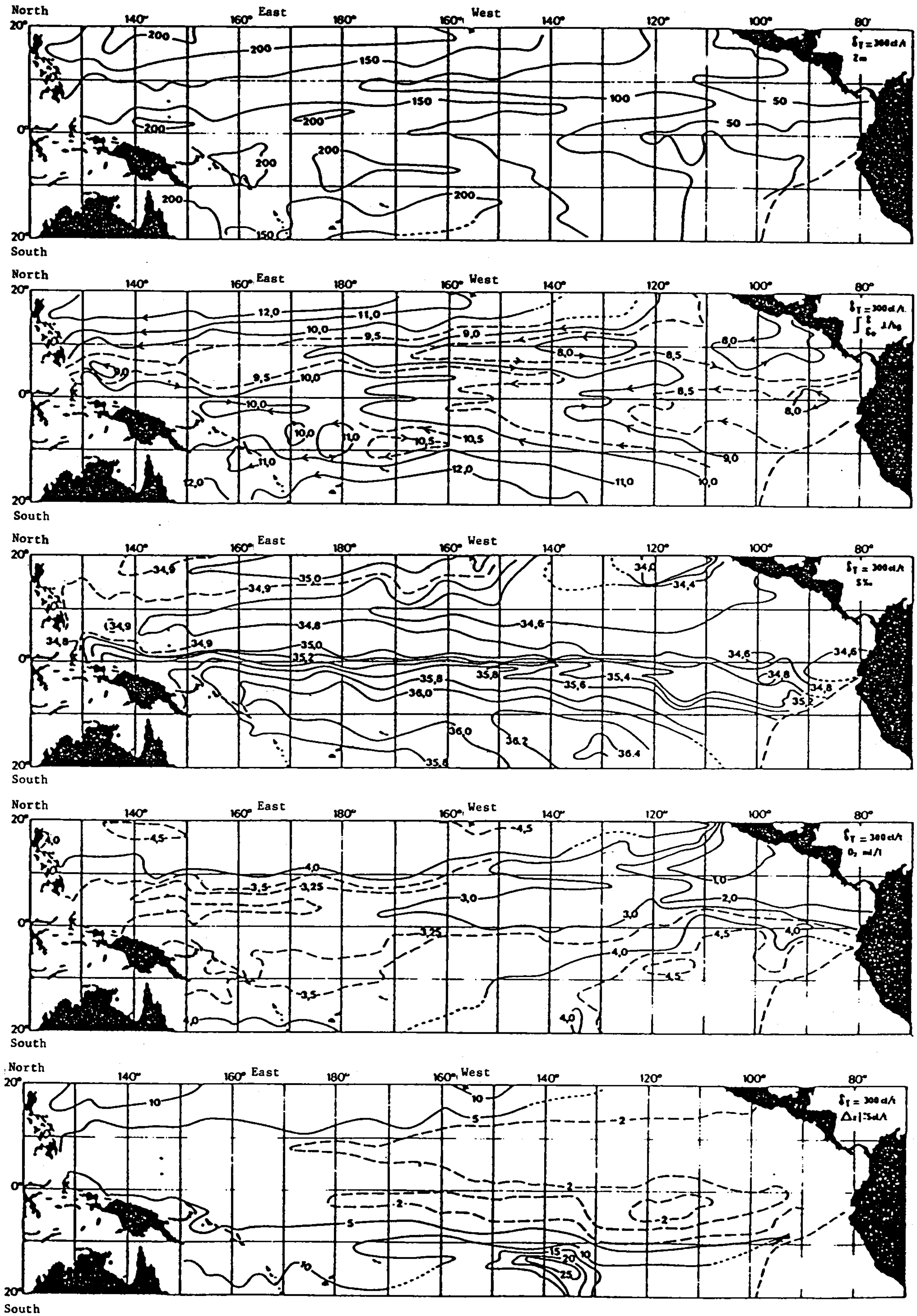

North
South
East
West
140°
160°
180°
120°
100°
80°
20°
0°
200
150
100
50
12,0
11,0
10,0
9,5
9,0
8,0
8,5
10,5
34,6
34,9
35,0
34,8
35,2
35,8
36,0
35,6
35,4
36,2
36,4
4,5
4,0
3,5
3,25
3,0
2,0
1,0
10
5
2
15
20
25

FIG. 18

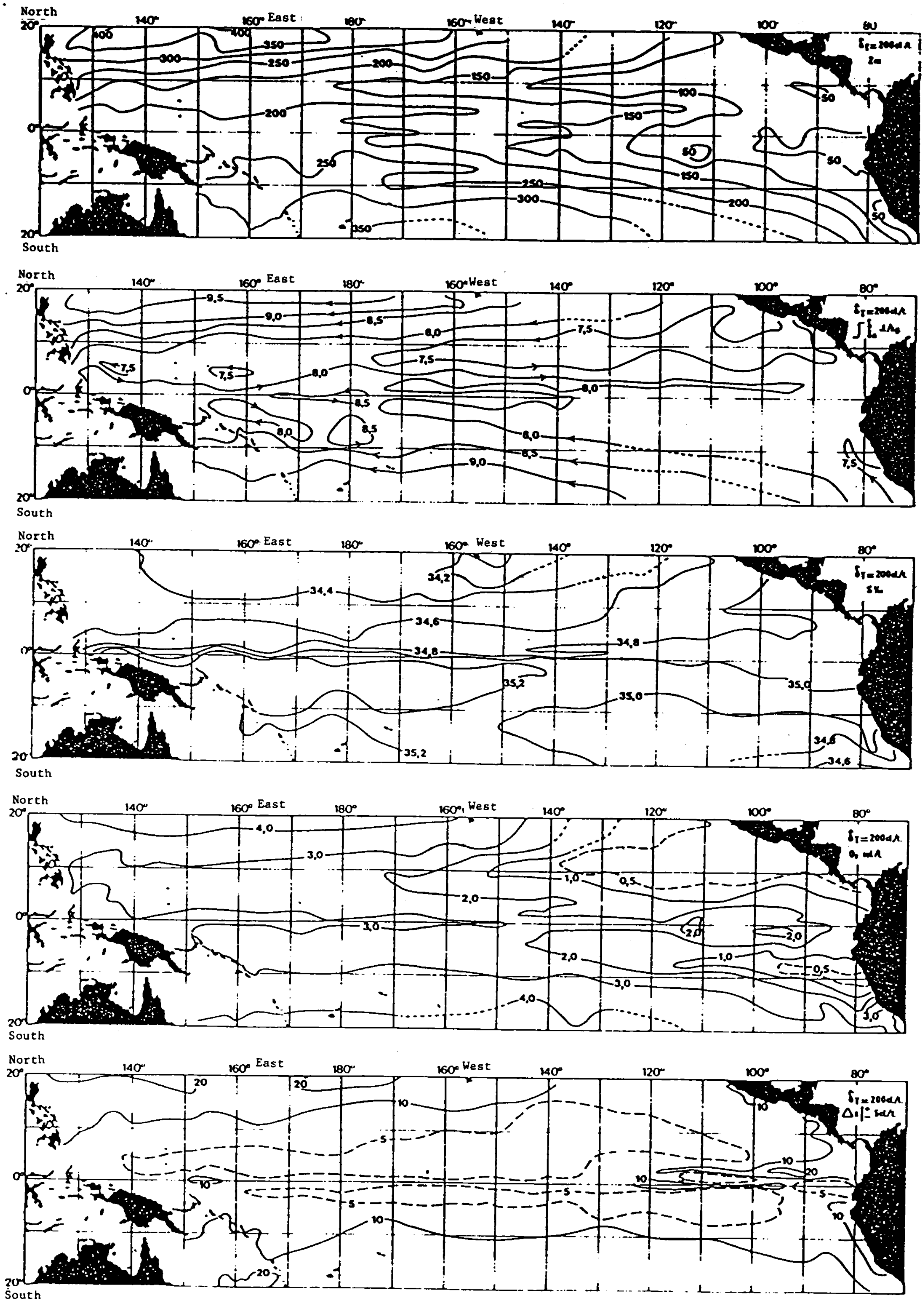

North
20°
140°
160° East
180°
160° West
140°
120°
100°
80°
0°
20°
South
400
350
300
250
200
150
100
50
9,5
9,0
8,5
8,0
7,5
34,2
34,4
34,6
34,8
35,0
35,2
4,0
3,0
2,0
1,0
0,5
20
10
5

FIG. 19

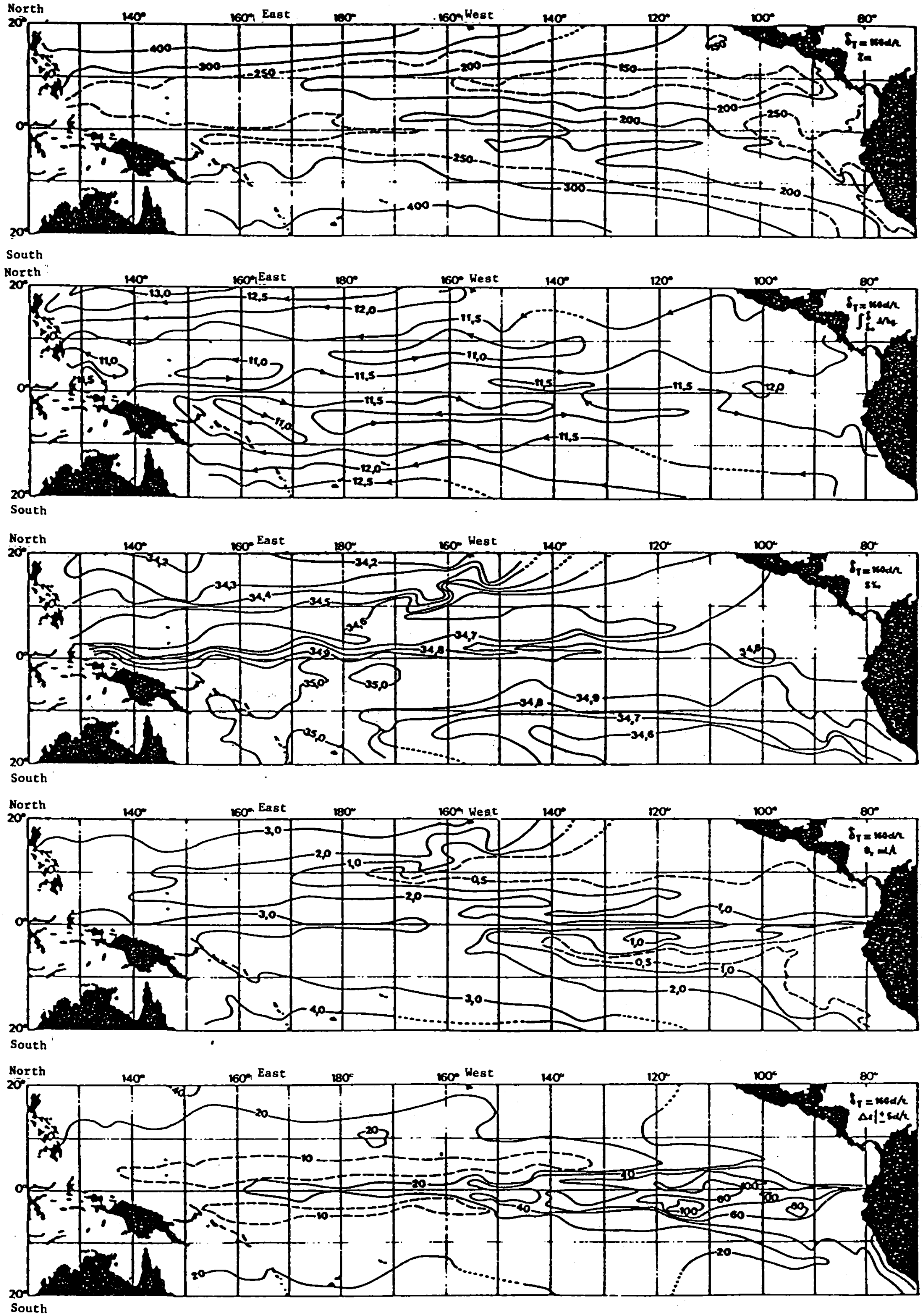
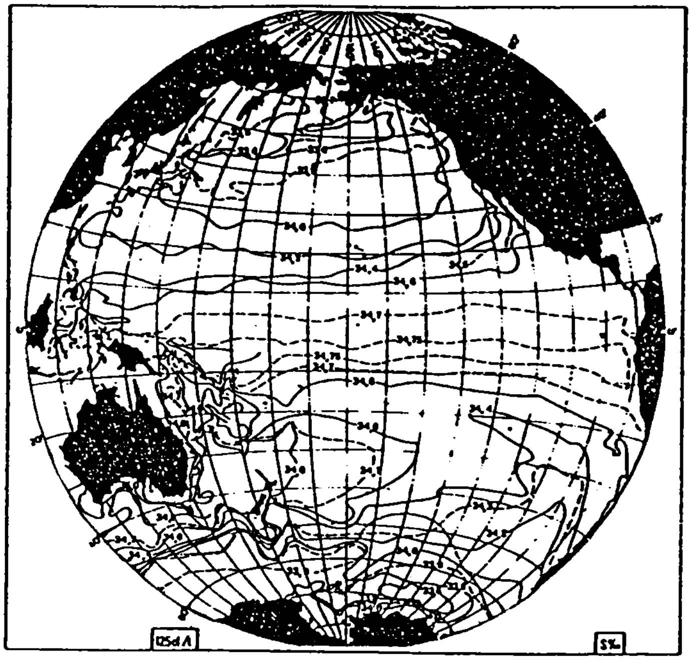
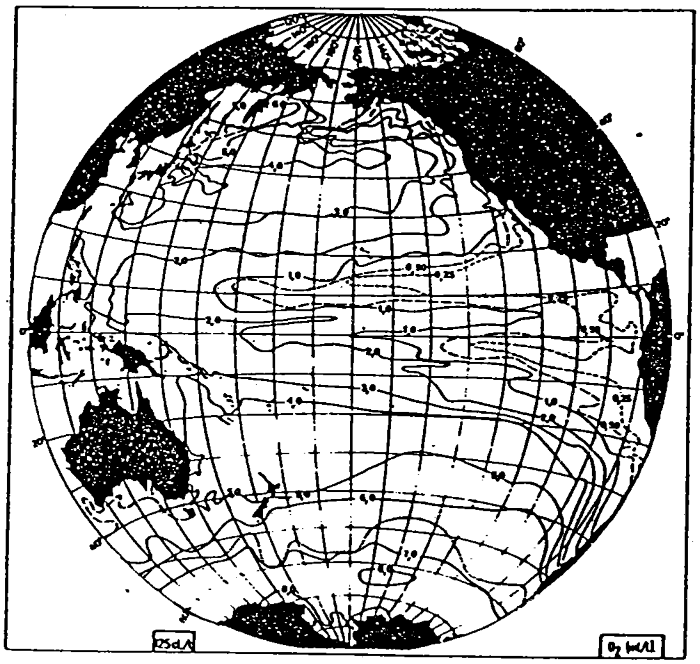
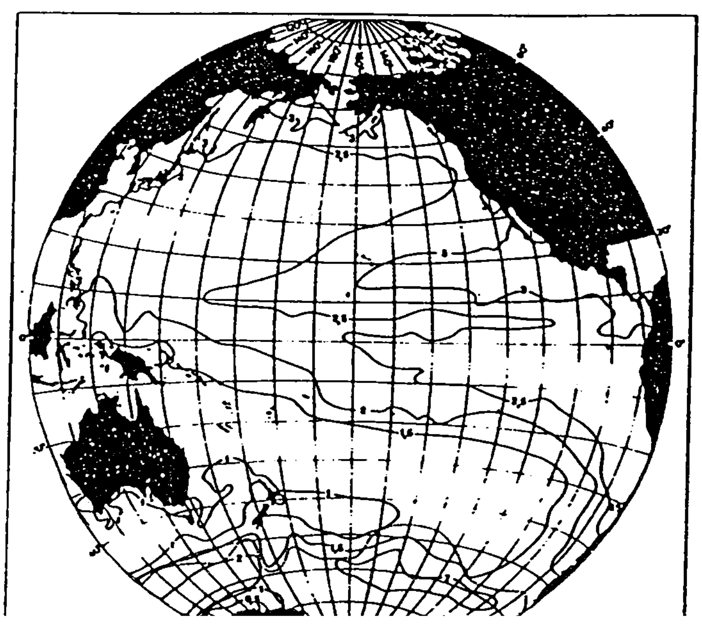

North
South
East
West
140°
160°
180°
120°
100°
80°
20°
0°
400
300
250
200
150
13,0
12,5
12,0
11,5
11,0
34,2
34,3
34,4
34,5
34,6
34,7
34,8
34,9
35,0
3,0
2,0
1,0
0,5
4,0
20
10
40
60
80
100

FIG. 20


125d A
S‰


125d A
O2 (ml/l)



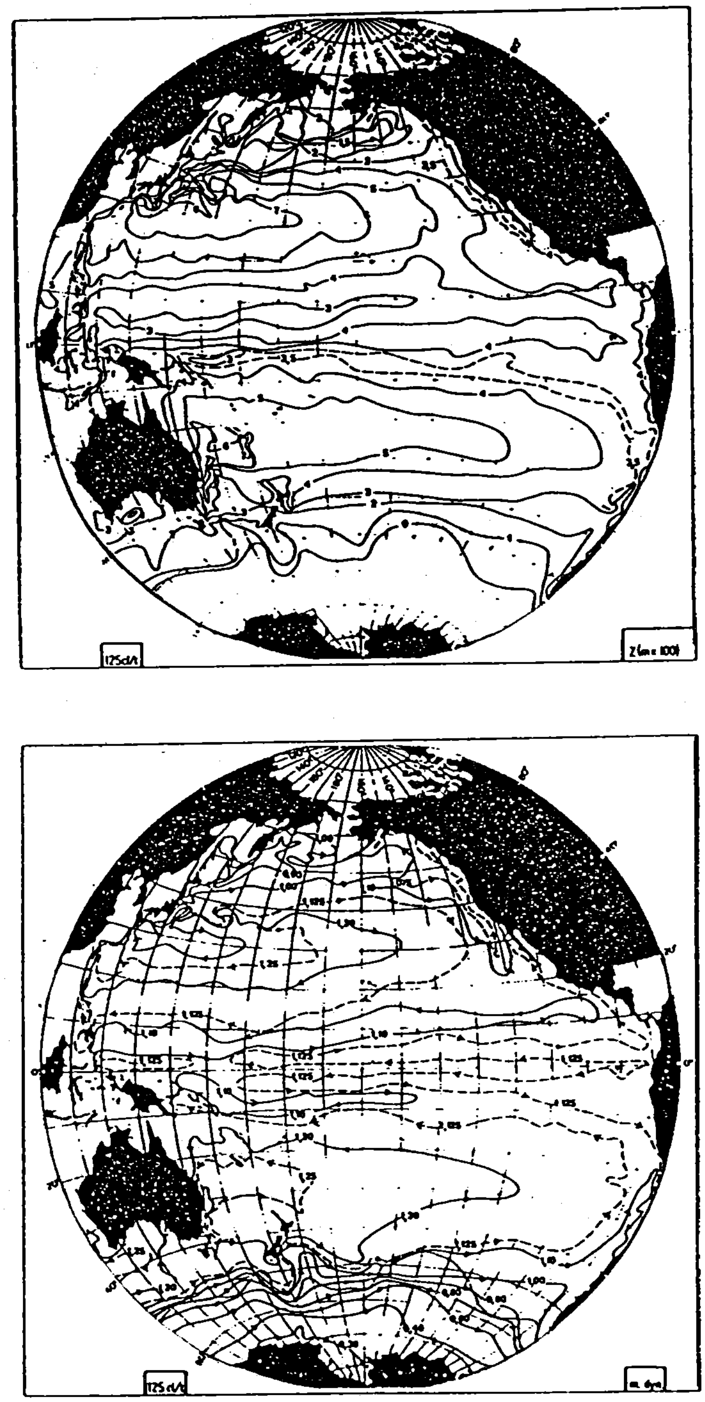
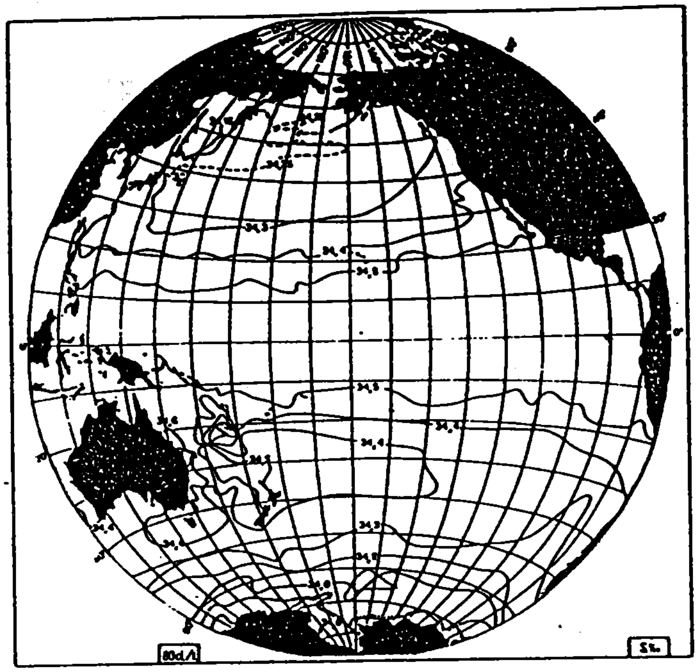
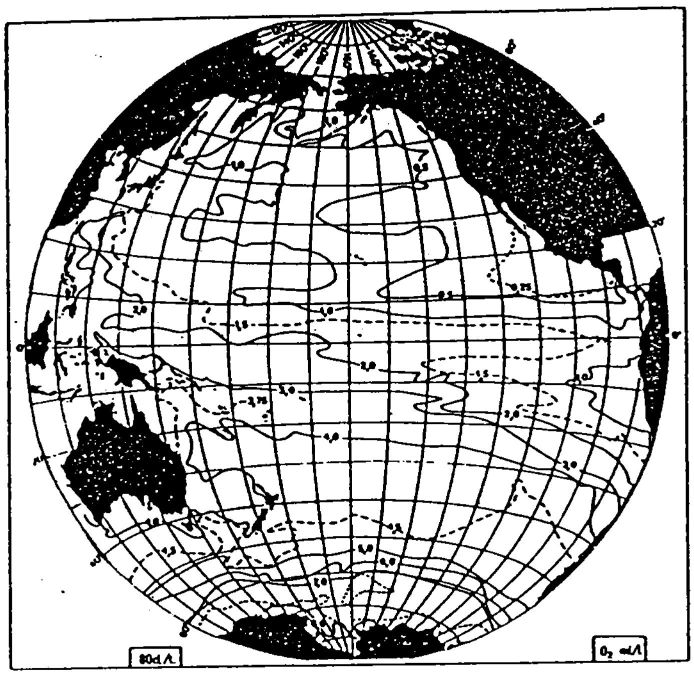
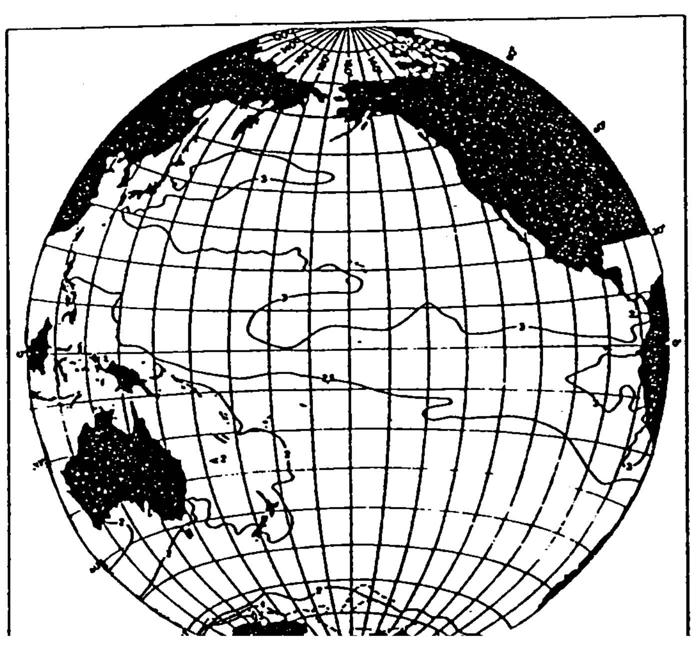


FIG. 22


80d/l
S‰


80d/l
O_2 ml/l



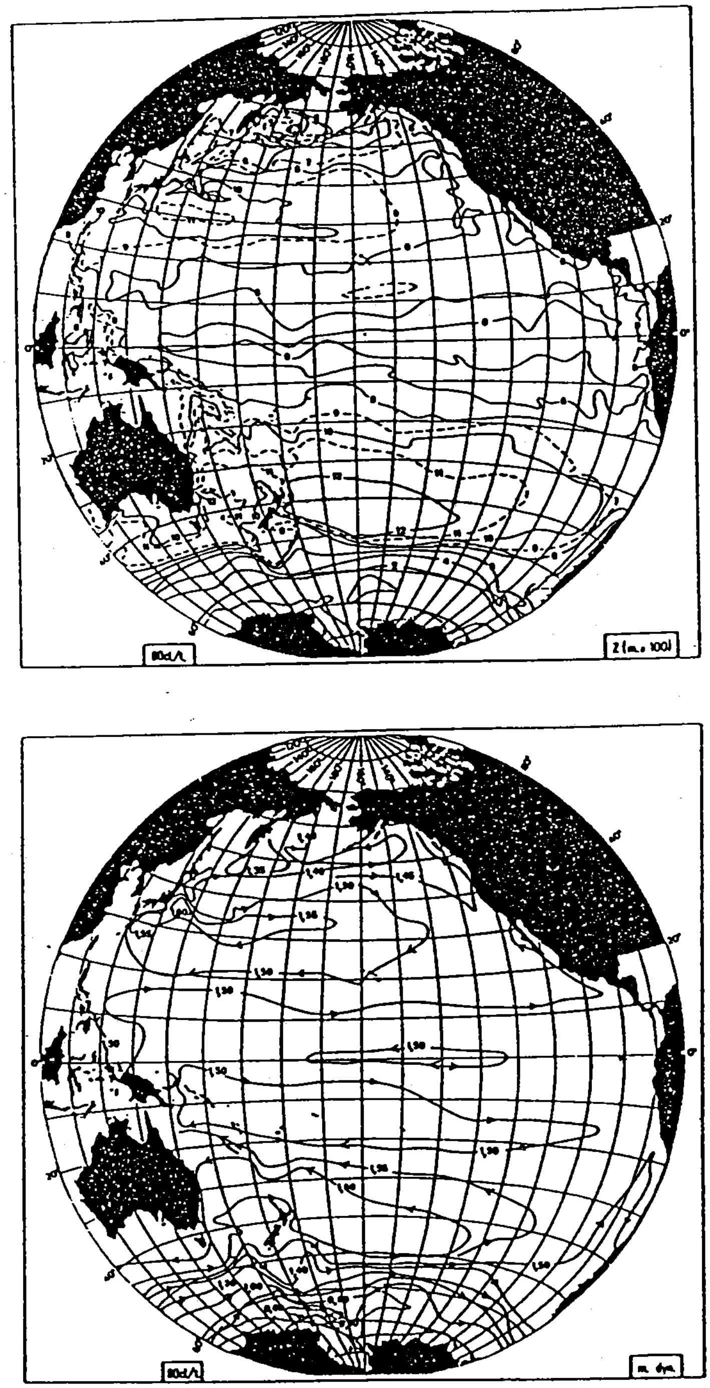


FIG. 24

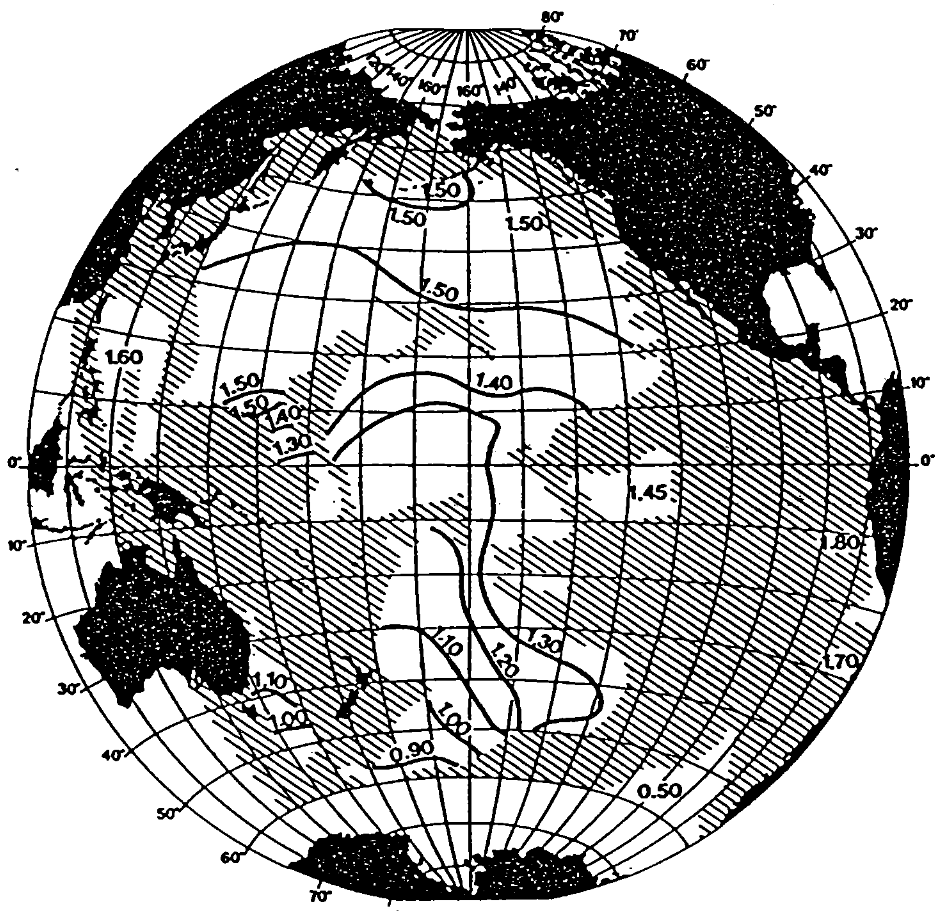
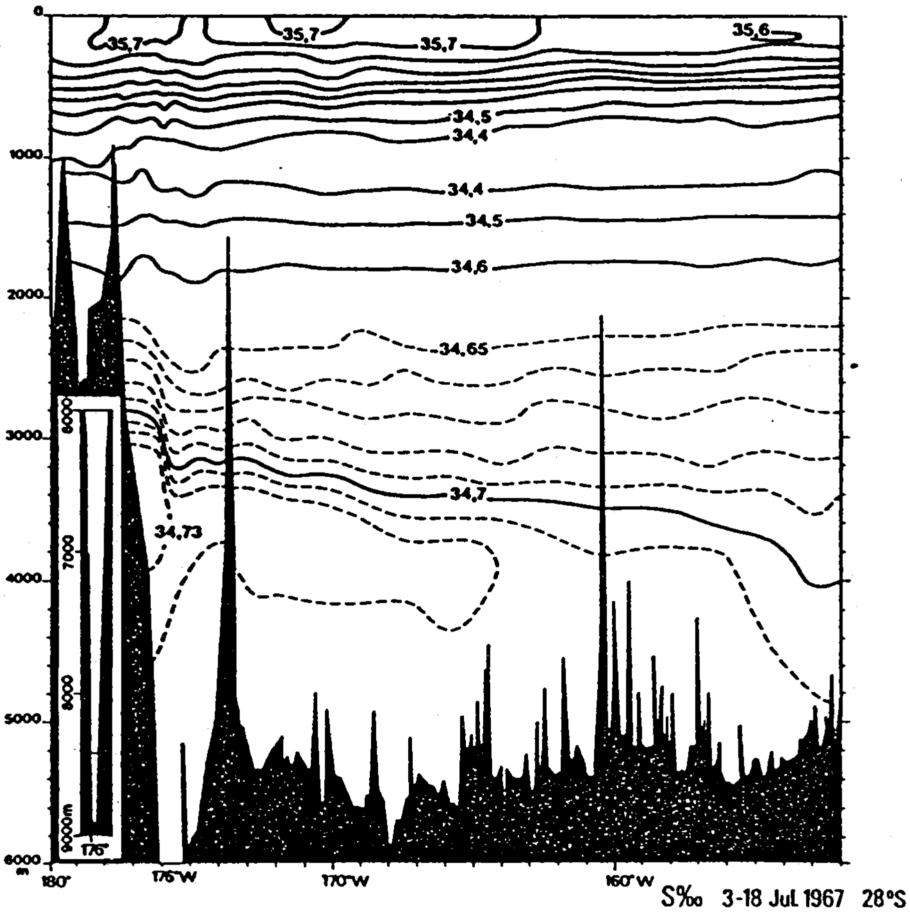
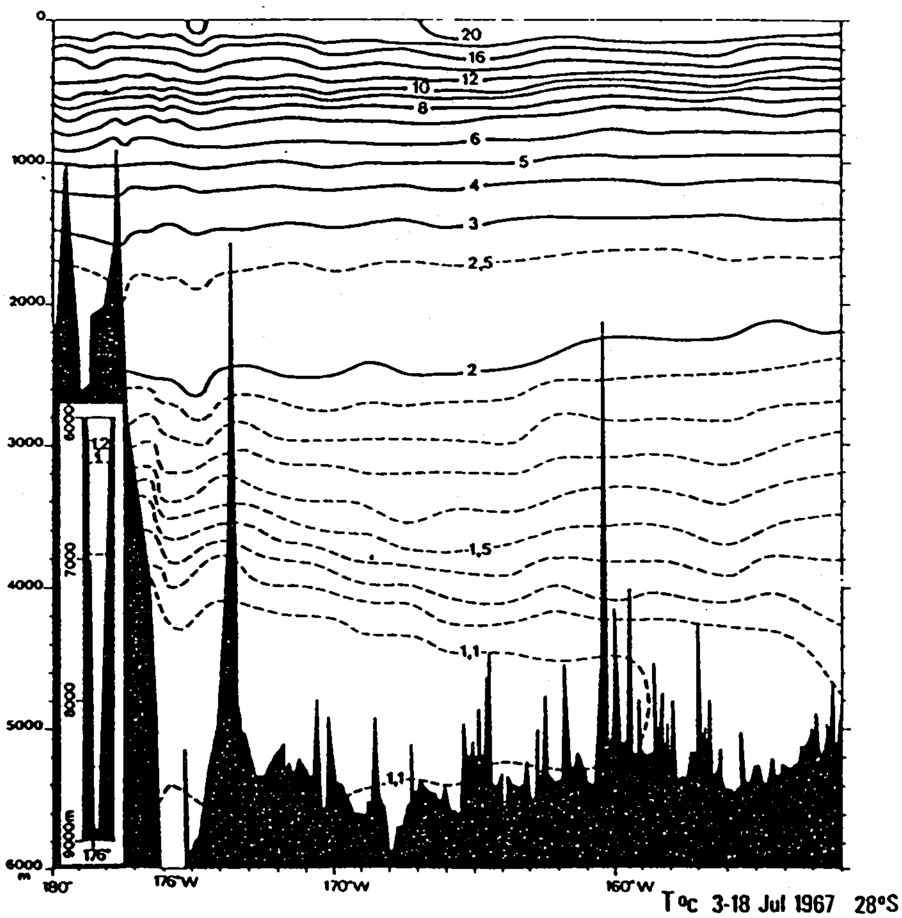

80°
70°
60°
50°
40°
30°
20°
10°
0°
1.50
1.50
1.50
1.50
1.60
1.50
1.50
1.40
1.30
1.40
1.45
1.80
1.10
1.20
1.30
1.70
1.10
1.00
1.00
0.90
0.50

FIG. 25


35,7
35,7
35,7
35,6
34,5
34,4
34,4
34,5
34,6
34,65
34,7
34,73
0
1000
2000
3000
4000
5000
6000
m
8000
7000
8000
9000m
176°
180°
176°W
170°W
160°W
S‰ 3-18 Jul 1967 28°S


20
16
12
10
8
6
5
4
3
2,5
2
1,2
1,1
1,5
1,1
1,1
0
1000
2000
3000
4000
5000
6000
m
8000
7000
8000
9000m
176°
180°
176°W
170°W
160°W
T°C 3-18 Jul 1967 28°S

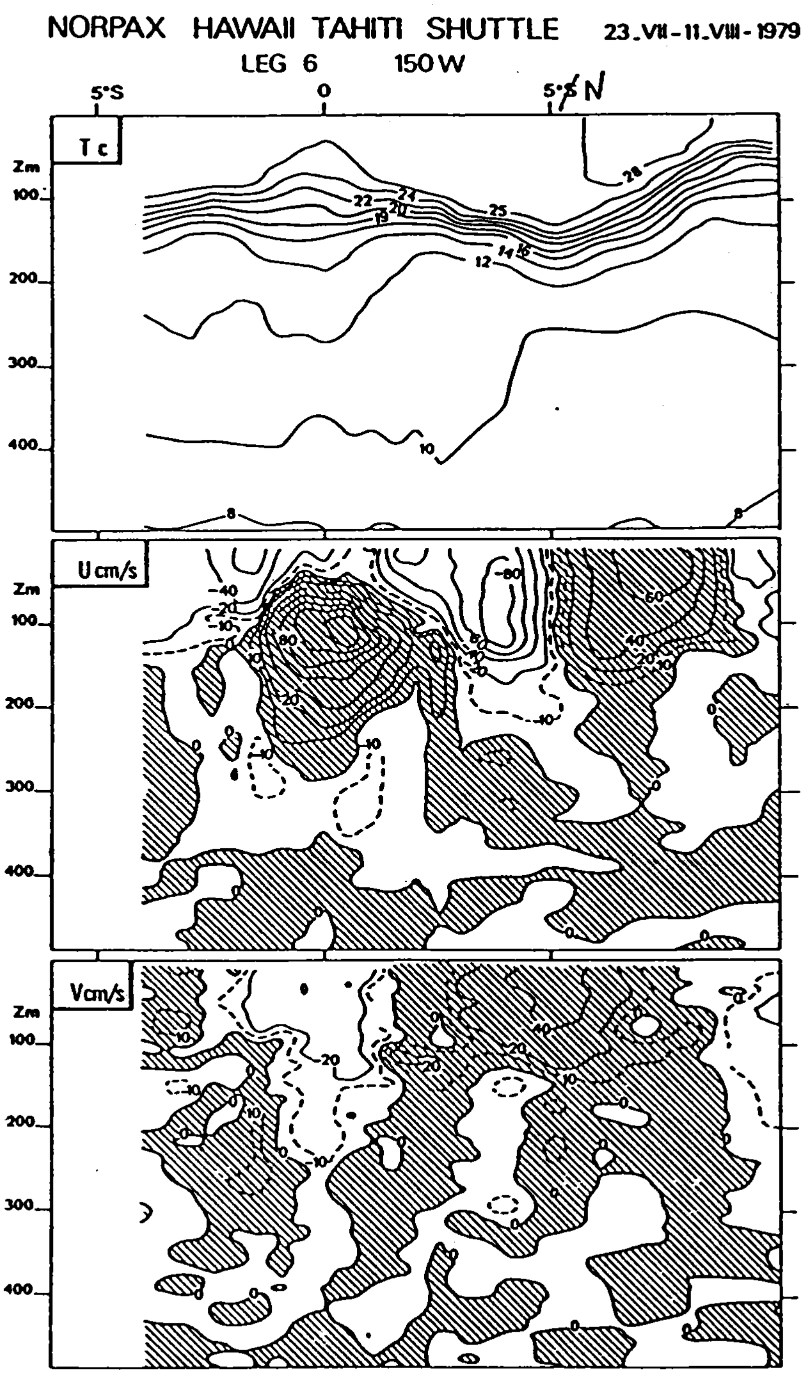
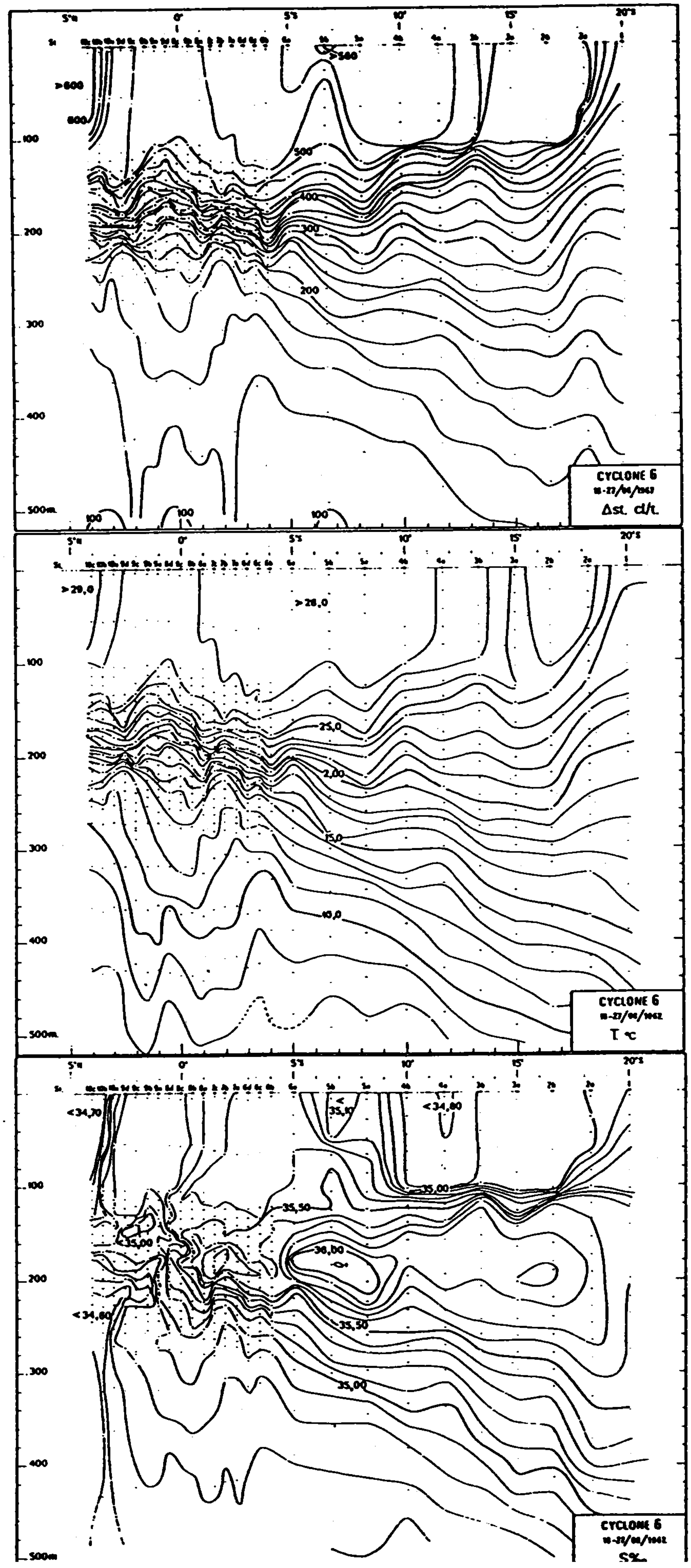
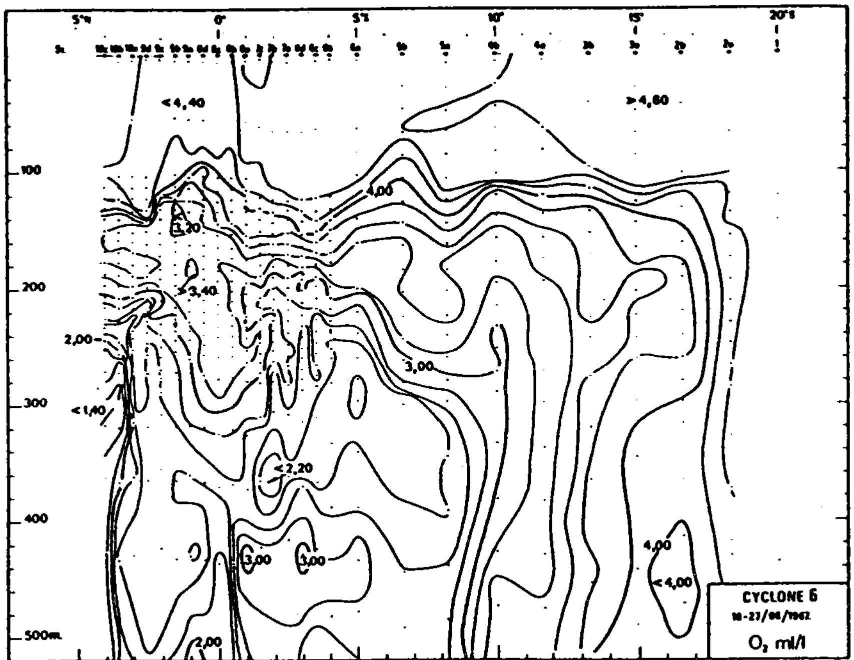
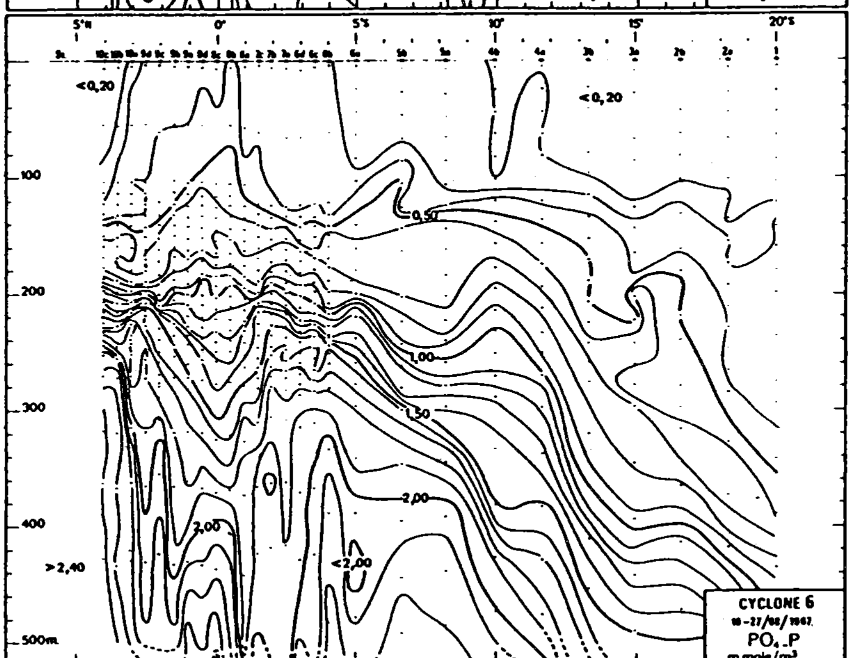
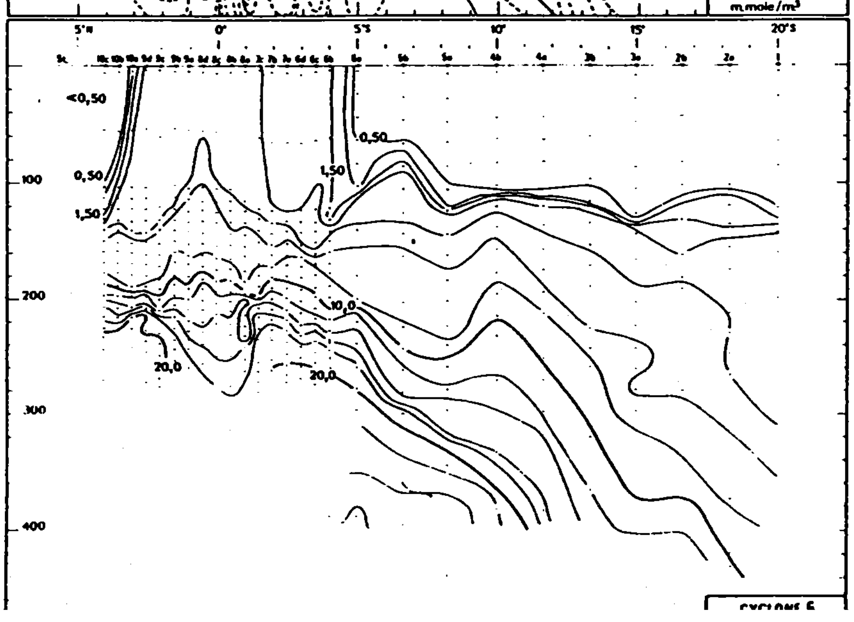

NORPAX HAWAII TAHITI SHUTTLE 23.VII-11.VIII-1979
LEG 6 150W
5°S
0
5°N
Tc
Zm
100
200
300
400
Ucm/s
Vcm/s

FIG. 27


5°N
0°
5°S
10°
15°
20°S
>600
600
500
400
300
200
100
100
200
300
400
500m
CYCLONE 6
Δst. cl/t.
>29,0
>28,0
25,0
20,0
15,0
10,0
CYCLONE 6
T °C
<34,70
35,10
<34,80
35,00
35,50
35,00
36,00
<34,80
35,50
35,00
CYCLONE 6
S‰


5°N
0°
5°S
10°
15°
20°S
<4,40
>4,80
100
4,00
3,20
200
>3,40
2,00
3,00
300
<1,40
<2,20
400
3,00
3,00
4,00
<4,00
500m
2,00
CYCLONE 6
O2 ml/l


5°N
0°
5°S
10°
15°
20°S
<0,20
<0,20
100
0,50
200
1,00
300
1,50
2,00
400
2,00
>2,40
<2,00
500m
CYCLONE 6
PO4-P
m.mole/m³


5°N
0°
5°S
10°
15°
20°S
<0,50
0,50
0,50
1,50
100
1,50
200
10,0
20,0
20,0
300
400

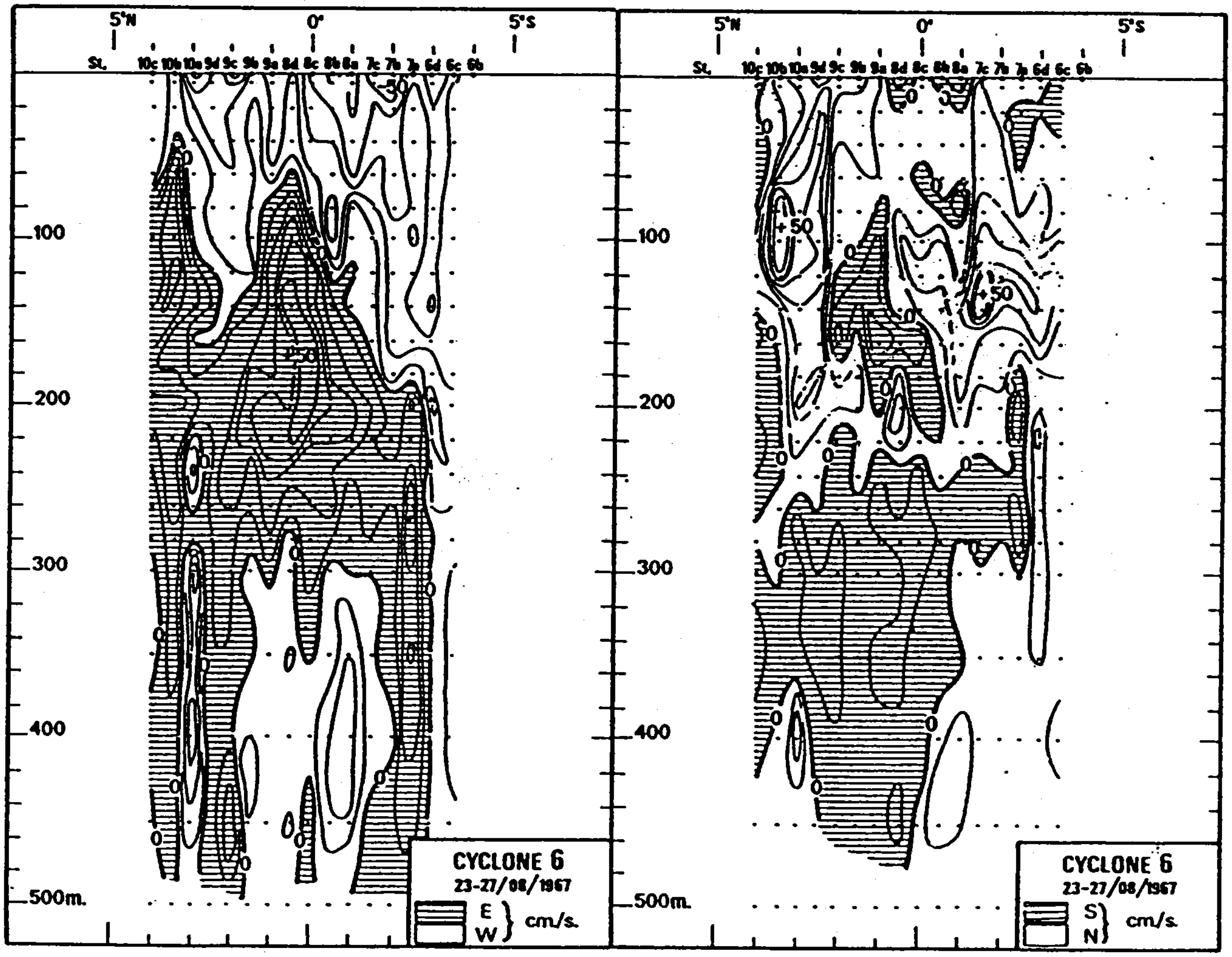
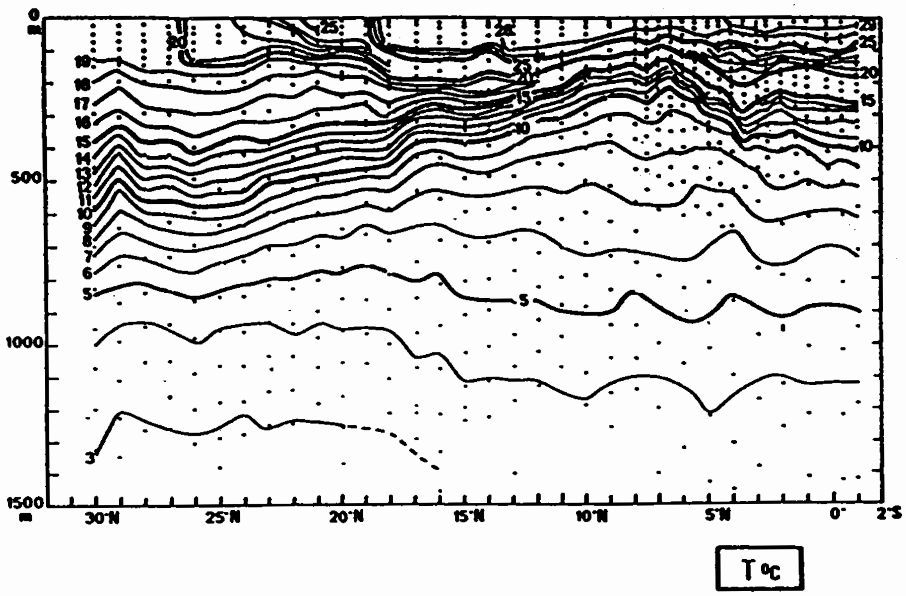
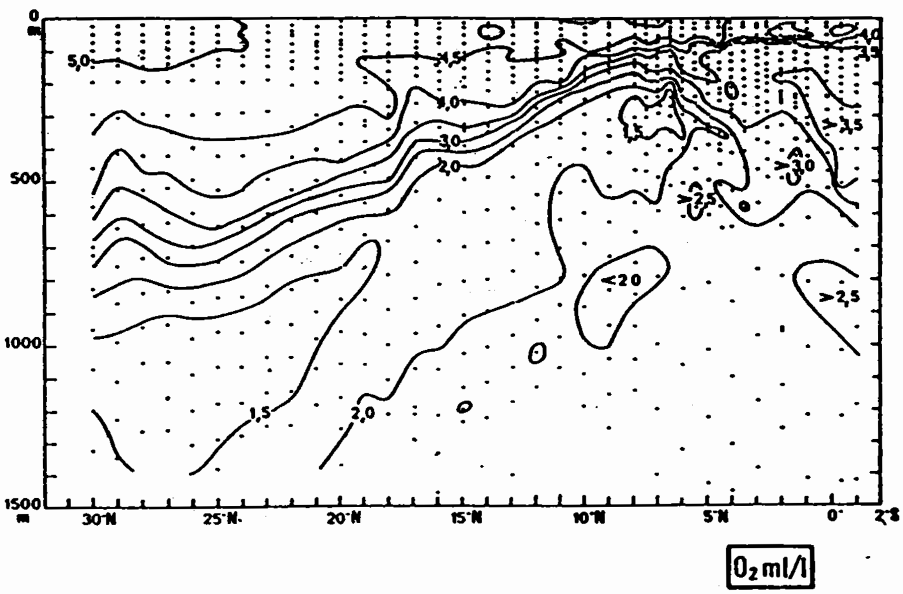
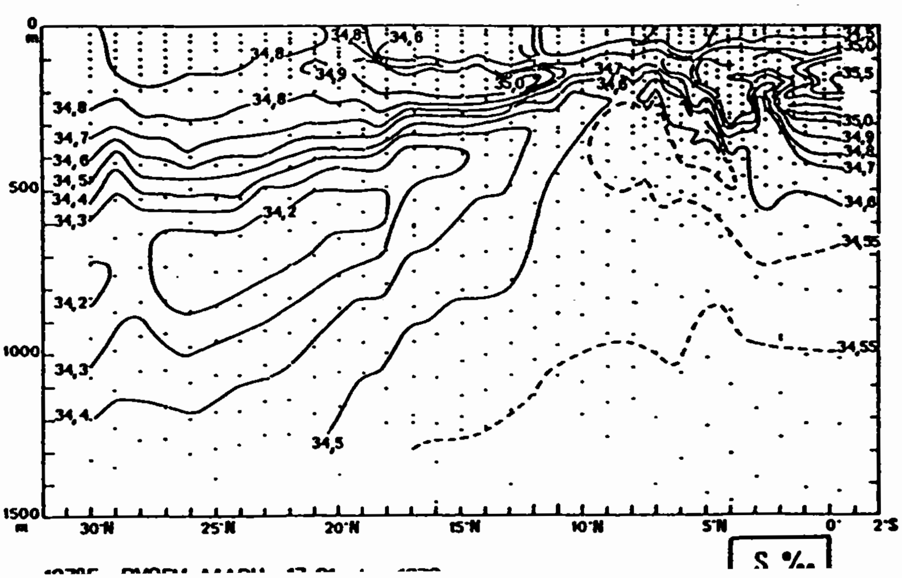
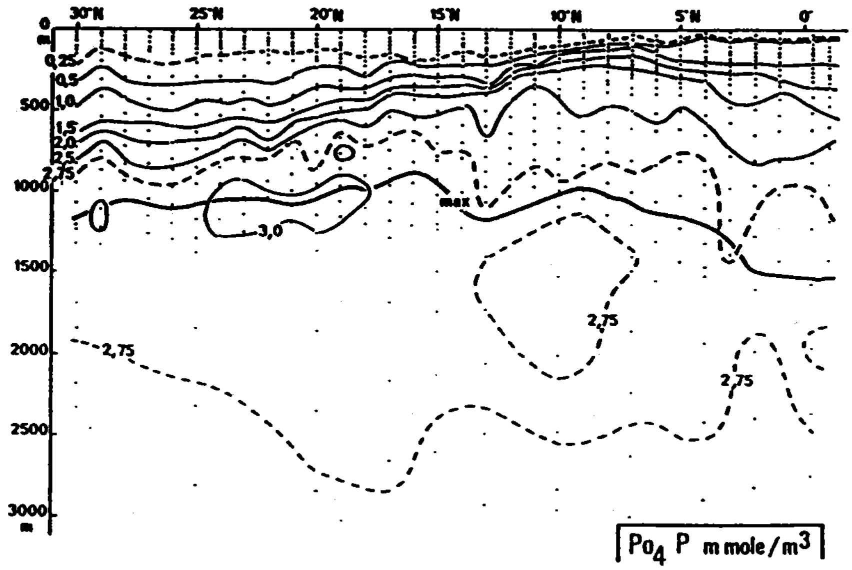

5°N
0°
5°S
St.
10c 10b 10a 9d 9c 9b 9a 8d 8c 8b 8a 7c 7b 7a 6d 6c 6b
100
200
300
400
500m.
CYCLONE 6
23-27/08/1967
E
W
cm/s.
S
N
+50
-50

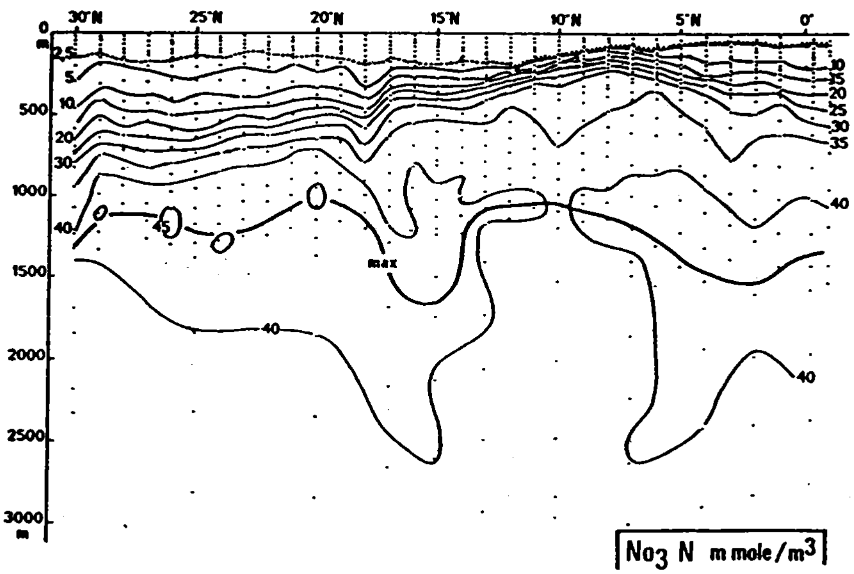
FIG. 30

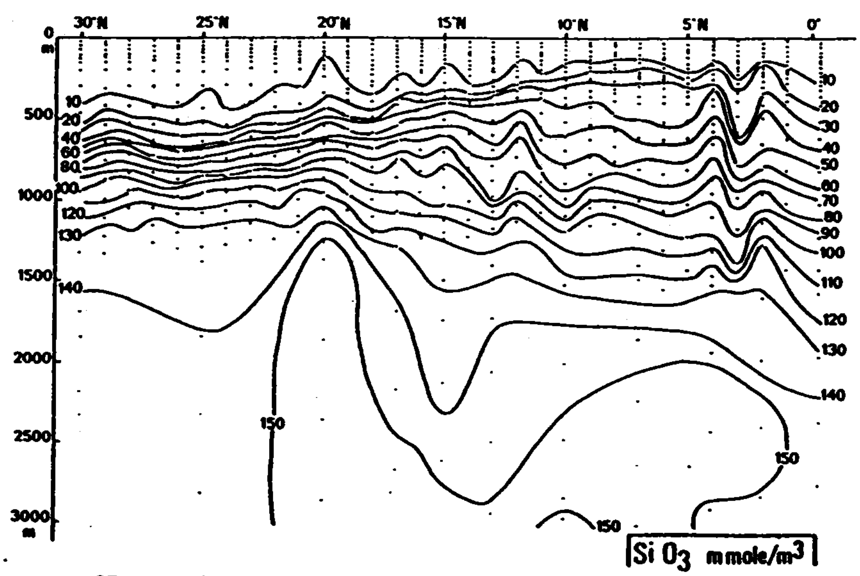

0 m
500
1000
1500 m
30°N
25°N
20°N
15°N
10°N
5°N
0°
2°S
T °C


0 m
500
1000
1500 m
30°N
25°N
20°N
15°N
10°N
5°N
0°
2°S
O2 ml/l


0 m
500
1000
1500 m
30°N
25°N
20°N
15°N
10°N
5°N
0°
2°S
S ‰







137°E RYOFU MARU 17-31 Jan 1970

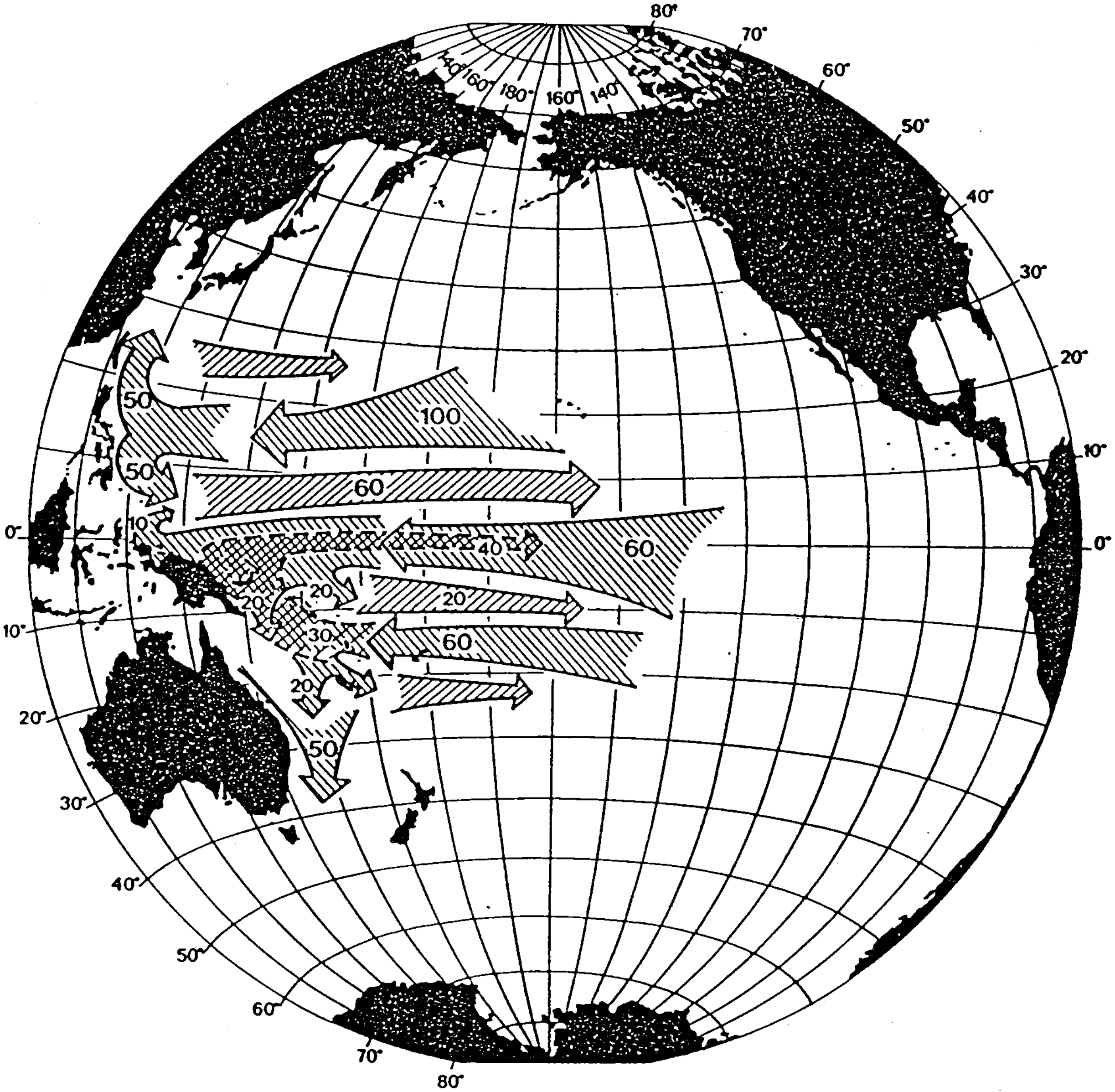

80°
70°
60°
50°
40°
30°
20°
10°
0°
140°
160°
180°
160°
140°
100
60
60
50
50
10
40
20
20
20
60
30
20
50
0°
10°
20°
30°
40°
50°
60°
70°
80°

FIG. 33

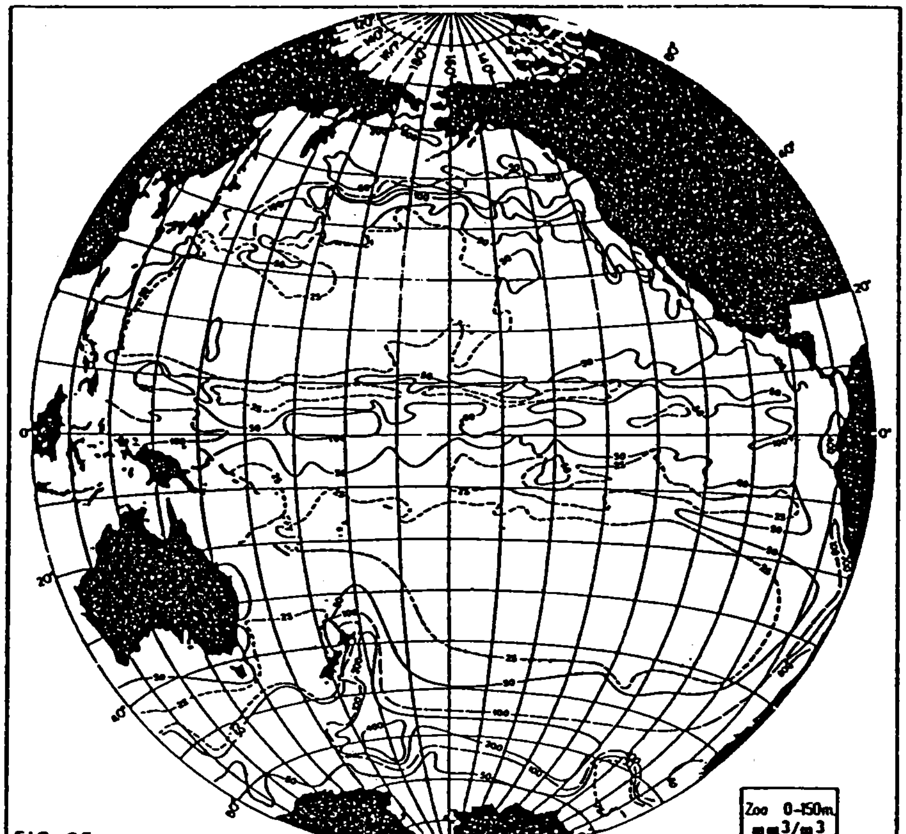


FIG. 35

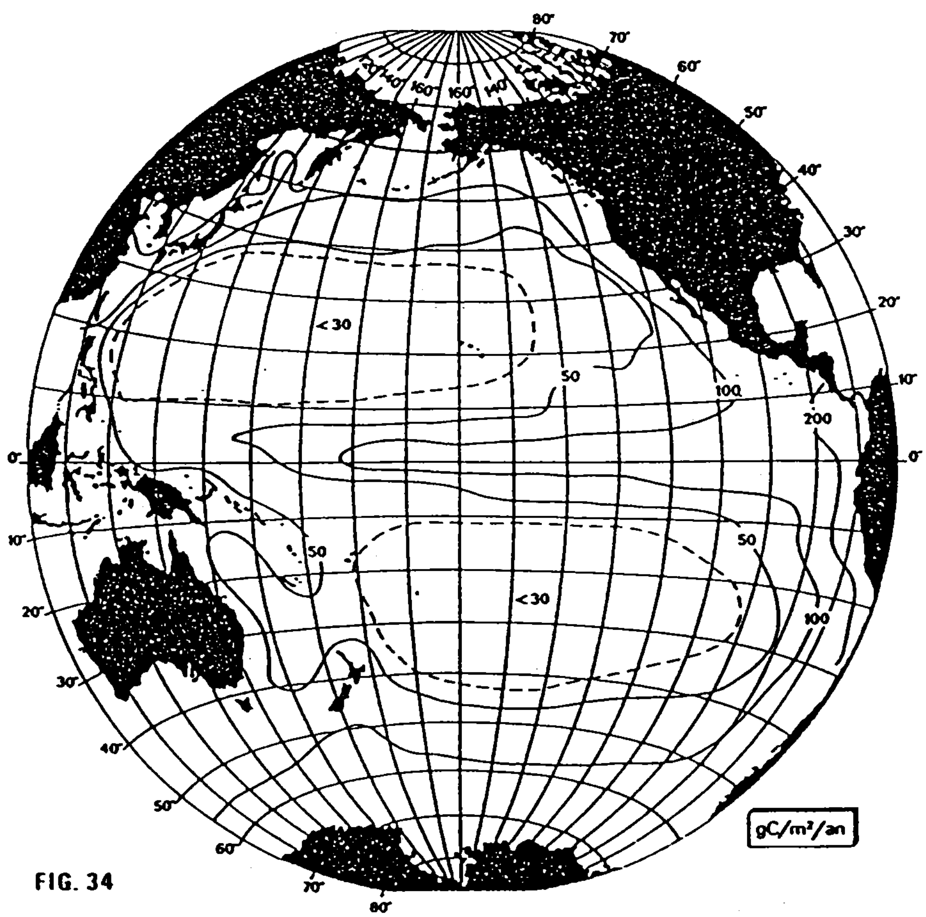


FIG. 34

PUBLICATIONS IN THE UNEP REGIONAL SEAS REPORTS AND STUDIES SERIES

No. 1 UNEP: Achievements and planned development of UNEP's Regional Seas Programme and comparable programmes sponsored by other bodies. (1982)

No. 2 UNIDO/UNEP: Survey of marine pollutants from industrial sources in the West and Central African region. (1982)

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No. 15 UNEP: Guidelines and principles for the preparation and implementation of comprehensive action plans for the protection and development of marine and coastal areas of regional seas. (1982)

No. 16 GESAMP: The health of the oceans. (1982)

No. 17 UNEP: Regional Seas Programme: Legislative authority. (1985)

No. 18 UNEP: Regional Seas Programme: Workplan. (1982)

No. 19 Rev. 2. UNEP: UNEP Oceans Programme: Compendium of projects. (1985)

No. 20 CPPS/UNEP: Action Plan for the protection of the marine environment and coastal areas of the South-East Pacific. (1983)

No. 21 CPPS/UNEP:Sources, levels and effects of marine pollution in the South-East Pacific. (1983) (In Spanish only)

No. 22 Rev. 2. UNEP: Regional Seas Programme in Latin America and Wider Caribbean. (1985)

No. 23 FAO/UNESCO/IOC/WHO/WMO/IAEA/UNEP: Co-ordinated Mediterranean Pollution Monitoring and Research Programme (MED POL) - Phase I: Programme Description. (1983)

No. 24 UNEP: Action Plan for the protection and development of the marine and coastal areas of the East Asian region. (1983)

No. 25 UNEP: Marine pollution. (1983)

No. 26 UNEP: Action Plan for the Caribbean environment programme. (1983)

No. 27 UNEP: Action Plan for the protection and development of the marine environment and coastal areas of the West and Central African region. (1983)

No. 28 UNEP: Long-term programme for pollution monitoring and research in the Mediterranean (MED POL) - Phase II. (1983)

No. 29 SPC/SPEC/ESCAP/UNEP: Action Plan for managing the natural resources and environment of the South Pacific region. (1983)

No. 30 UNDIESA/UNEP: Ocean energy potential of the West and Central African region. (1983)

No. 31 A. L. DAHL and I. L. BAUMGART: The state of the environment in the South Pacific. (1983)

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No. 33 UNDIESA/UNEP: Onshore impact of offshore oil and natural gas development in the West and Central African region. (1984)

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No. 55/ Annex FAO/IUCN/IWC/UNEP: Marine mammals: global plan of action. (1985)

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No. 60 IUCN/UNEP: Management and conservation of renewable marine resources in the Indian Ocean region: Overview. (1985)

No. 61 UNEP: Action Plan for the protection, management and development of the marine and coastal environment of the Eastern African region. (1985)

No. 62 IUCN/UNEP: Management and conservation of renewable marine resources in the South Asian Seas region. (1985)

No. 63 IUCN/UNEP: Management and conservation of renewable marine resources in the Kuwait Action Plan region. (1985)

No. 64 IUCN/UNEP: Management and conservation of renewable marine resources in the Red Sea and Gulf of Aden region. (1985)

No. 65 IUCN/UNEP: Management and conservation of renewable marine resources in the East Asian Seas region. (1985)

No. 66 IUCN/UNEP: Management and conservation of renewable marine resources in the Eastern African region. (1985)

No. 67 UN/UNEP: Coastal erosion in West and Central Africa. (1985)

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No. 75 UNEP: Environmental problems of the marine and coastal area of Bangladesh: National Report (1986)

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No. 82 UNEP: Environmental problems of the South Asian Seas region: An overview. (1986)

No. 83 SPC/SPEC/ESCAP/UNEP: B. Wauthy: Physical ocean environment in the South Pacific Commission Area. (1986)