KONINKLIJK NEDERLANDS METEOROLOGISCH INSTITUUT

MEDEDELINGEN EN VERHANDELINGEN

No. 96

L. OTTO

OCEANOGRAPHY OF THE RÍA DE AROSA (N.W. SPAIN)

1975

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STAATSUITGEVERIJ / 'S-GRAVENHAGE

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VOORWOORD

Het in de laatste jaren zo sterk beklemtoonde belang van het mariene milieu en van de onderlinge samenhang tussen de verschillende fysische, chemische, geologische en biologische processen in zee heeft geleid tot intensivering van het oceanografisch onderzoek. Dit onderzoek vereist veelal een gezamenlijke of multi-disciplinaire aanpak. Het fysisch onderzoek is hierbij vaak essentieel om tot een kwantitatieve benadering te komen.

Reeds in de jaren 1954 tot 1957 werd daarom door het KNMI medewerking verleend aan een gecombineerd onderzoek van het Eems-estuarium. De toen uitgevoerde studie en met name het uitwisselingsmodel dat daarbij door de oceanografische onderzoekers van het KNMI werd opgesteld blijkt van veel betekenis te zijn geweest bij de studie van de vervuilingsproblematiek in dit gebied.

Deze factoren hebben de toenmalige hoofddirecteur van het KNMI, Ir. C. J. Warners, er toe geleid om ook medewerking te verlenen aan een onderzoek dat in 1962-1964 door de Rijksuniversiteit van Leiden en het Rijksmuseum van Natuurlijke Historie in samenwerking met enkele andere instellingen, waaronder het Nederlands Instituut voor Onderzoek der Zee, werd uitgevoerd in de Ría de Arosa in N.W. Spanje. Hierbij speelde vooral ook de overweging een rol dat de ervaring bij een dergelijk onderzoek opgedaan van nut zou zijn bij het onderzoek van de Noordzee en de Nederlandse estuaria.

De onderhavige publicatie geeft een verslag van de uitkomsten van het werk verricht in de Ría de Arosa. Het bleek ook hier mogelijk een uitwisselingsmodel op te stellen dat gezien kan worden als een uitbreiding van het hiervoor genoemde Eems-model. Twee elementen zijn hier speciaal te vermelden. Ten eerste is de waarde van het model voor een meer kwantitatieve benadering van de chemische processen gebleken, en ten tweede blijkt in het betrokken gebied de nauwe samenhang met atmosferische processen en met name met de wind wel heel markant te zijn. Hoewel de koppeling tussen atmosfeer en oceaan op zichzelf geen recente ontdekking is, wordt juist de laatste tijd de vaak overwegende invloed van deze factor onderkend.

Bij het huidige onderzoek van de Noordzee, wordt eveneens gestreefd naar de ontwikkeling van bruikbare rekenmodellen. Om tot een oplossing te komen zijn zowel bij het veldwerk als bij het theoretisch onderzoek geavanceerde technieken nodig. Desondanks is het te verwachten dat het inzicht, bij de onderhavige studie met betrekkelijk eenvoudige middelen verworven, ook van belang zal zijn voor het verdere onderzoek in de Noordzee.

Deze studie werd door de Faculteit der Wiskunde en Natuurwetenschappen van de Rijksuniversiteit te Utrecht aanvaard als dissertatie.

De hoofddirecteur van het

Koninklijk Nederlands Meteorologisch Instituut

M. W. F. SCHREGARDUS

PREFACE

The emphasis of late years on the importance of the marine environment and on the interrelations between the different physical, chemical and biological processes in the sea has resulted in increased oceanographic research. This research often needs a combined or a multidisciplinary approach. Physical investigations are on many occasions essential in order to obtain quantitative estimates.

As early as 1954-1957 the KNMI (Royal Netherlands Meteorological Institute) took part in combined investigations in the Ems estuary. The work carried out on that occasion and in particular the model describing the water exchange that was developed by oceanographers from KNMI turned out to be important for the study of pollution in that area.

For these reasons the director-in-chief of KNMI at that time, Ir. C. J. Warners lent his support to a programme of investigations carried out in the Ría de Arosa (N.W. Spain) by the State University of Leiden and the 'Rijksmuseum van Natuurlijke Historie' together with some other institutes, including the Netherlands Institute for Sea Research. A special point of consideration was the expectation that the experience obtained during such an investigation would be useful in the study of the North Sea and the estuaries in the Netherlands.

The present publication reports on the results of the investigations made in the Ría de Arosa. Here it was also found possible to develop a model for the water exchange that can be considered as an extension of that for the Ems. There are two points that deserve special attention. In the first place the value of the model for a more quantitative approach of the chemical processes is apparent. In the second place in this area the close connection with the atmospheric processes and especially with the wind is marked. The interaction between atmosphere and ocean, although no new discovery, is now recognized as often having a predominating influence.

The present studies of the North Sea are also directed towards the development of practical models. To find a solution highly developed techniques both for observations and for theoretical work are needed. Nevertheless, it is expected that the insight obtained in the present study by comparatively simple means, will also be important for the further study of the North Sea.

This study has been accepted by the Faculty of Natural Sciences of the University of Utrecht as a thesis for the degree of doctor.

The Director in Chief of the

Royal Netherlands Meteorological Institute

M. W. F. SCHREGARDUS

ABSTRACT

In this study the oceanography of the Ría de Arosa is dealt with. The investigation is based on observations from three oceanographic campaigns during the summers of 1962, 1963 and 1964, and one campaign during the winter of 1964. The Ría de Arosa is situated in an area with a marked annual climatological variation. Of direct importance are the variation of the wind, which at sea during the summer is highly persistent from northerly directions, thus causing coastal upwelling from July till September, and the precipitation, which is high during the winter and low during the summer, thus giving a marked annual variation of the river discharge into the ria. Diurnal variations of the wind are notable in the coastal regions. In the inner part of the ria this leads to a modulation of the northerly winds in such a way that during the night strong northerly winds prevail, while during daytime southwesterly winds also occur regularly, although the northerly wind also frequently blows often attaining windforce 5 or more.

Estimates of the summer discharge into the ria from the largest river, the Rio Ulla, give values of about 11.5 m^3/s. The second largest river, the Rio Umia was estimated to have a discharge of about 2 m^3/s. The remaining run-off is negligibly small. Evaporation in the hinterland is high, but estimates of the evaporation over the sea show low values, because of the low temperature of the surface water.

The upwelling in summer brings water masses from below 200 m depth on to the shelf. These water masses have the temperature-salinity relationship typical for North Atlantic Central water. Sometimes water of higher salinities is observed on the shelf at sub-surface depths.

The Ría de Arosa has a complicated topography. The maximum depth is somewhat over 70 m. The inner parts have average depths of 11 m. Tidal motion is important (tidal range over 3 m at spring-tide), other types of sea level variations and waves are of secondary importance.

The structure of the water masses in summer shows a stratification. There is an inflow of cold, saline oceanic water (partially North Atlantic Central water) at greater depths and an outflow of warmer, less saline water at the surface. This circulation pattern is not only caused by density differences, but also by the predominant northerly winds. The circulation is accompanied by upwelling in the inner ria, which is the reason for the rather low surface temperatures observed. Details of this circulation are given to some extent. Special situations found in secondary bays are dealt with cursorily.

In winter the observations show a less important stratification. However, as the meteorological conditions during the winter campaign were not representative for the normal situation and as the river discharge was far below normal, no general conclusions can be drawn from these data.

Summer observations of suspended matter from the ria itself indicate the presence of a large fraction of organic material. Data from the rivers point to the presence of a turbidity maximum.

The maximum compensation depth is estimated to be between 25 and 40 m. In the surface layers concentrations of nutrients are low and the concentration of dissolved oxygen is high. In the deep layers the reversed situation was found, but extreme low oxygen values were observed nowhere (minimum 3.7 ml/l). There are, however, deviations from this rule at the surface in the upwelling regions, where high nutrient concentrations and undersaturation of oxygen were found.

In this study the observations are used to set up a model for the summer circulation of the inner and central parts of the ria. This model is consequently applied to the heat balance and the balance of nutrients and oxygen. The heat balance does not give fully satisfactory results in detail, but the overall picture is quite acceptable. The balance of nutrients enables to estimate a mean value of the productivity. The oxygen balance shows a deficit for the deep layers.

Finally the summer situation of the Ría de Arosa is compared with existing theories on estuarine circulation and mixing and with reported conditions in other embayments. Some points of correspondence with fjords and with the wind-driven circulation of the Gulf of Cariaco are noted. However, it appears that under normally prevailing conditions anoxic bottom waters that are found in these latter embayments are unlikely to occur in the Ría de Arosa.

CONTENTS

1. General introduction

1.1 Aim and scope of this study

The following study gives a description of the oceanographic situation in the Ría de Arosa in northwestern Spain with a twofold objective. In the first place a general description supplies information needed for the interpretation of the results of a number of investigations on the biology and geology of this embayment and its surroundings. In the second place this investigation may give a contribution to the oceanographic study of various types of estuaries, bays, fjords and other coastal areas with their diversified circulations and mixing processes.

It appeared impossible to draw a complete and unambiguous picture of the oceanographic situation of this embayment. Yet it is thought that in several respects the present investigation may attain the goals indicated.

1.2 The Rías Bajas of Galicia

The Ría de Arosa is the largest of the four 'Rías Bajas' of Galicia (N.W. Spain). Their geographical position is given in fig. 1.1. These Rías Bajas are four elongated bays that penetrate the west coast of Galicia in an approximately northeastern direction. From the north to the south we find the Ría de Muros y Noya, the Ría de Arosa, the Ría de Pontevedra and the Ría de Vigo. A great number of bays along the north and west coasts of Spain are also named 'Ría', but these four Rías Bajas clearly take a special position because of their size and their general form. Whether they can also be considered to be a special group among the other bays and estuaries of the Spanish Atlantic coast, from the point of view of physical oceanography, will be discussed at the end of this study.

In the geological and geographical literature the term 'ria' denotes a drowned river valley, and rias, in this sense of the word, have been described from different parts of the world. As regards the water properties and circulation of course they may be very different.

A discussion of the geological and geomorphological aspects of the Rías Bajas has been given by PANNEKOEK (1966 a, b, 1970). A detailed study of the geomorphology of Galicia, containing much information of general interest has been given in the publication by NONN (1966).

1.3 The Ría de Arosa, general geography

The Ría de Arosa is not only the largest, but is also the most irregularly shaped of the four Rías Bajas. Its surface is about 230 km^2, its length from the Isla Sálvora at the entrance to the mouth of the estuary of the Rio Ulla is about 25 km. The minimum

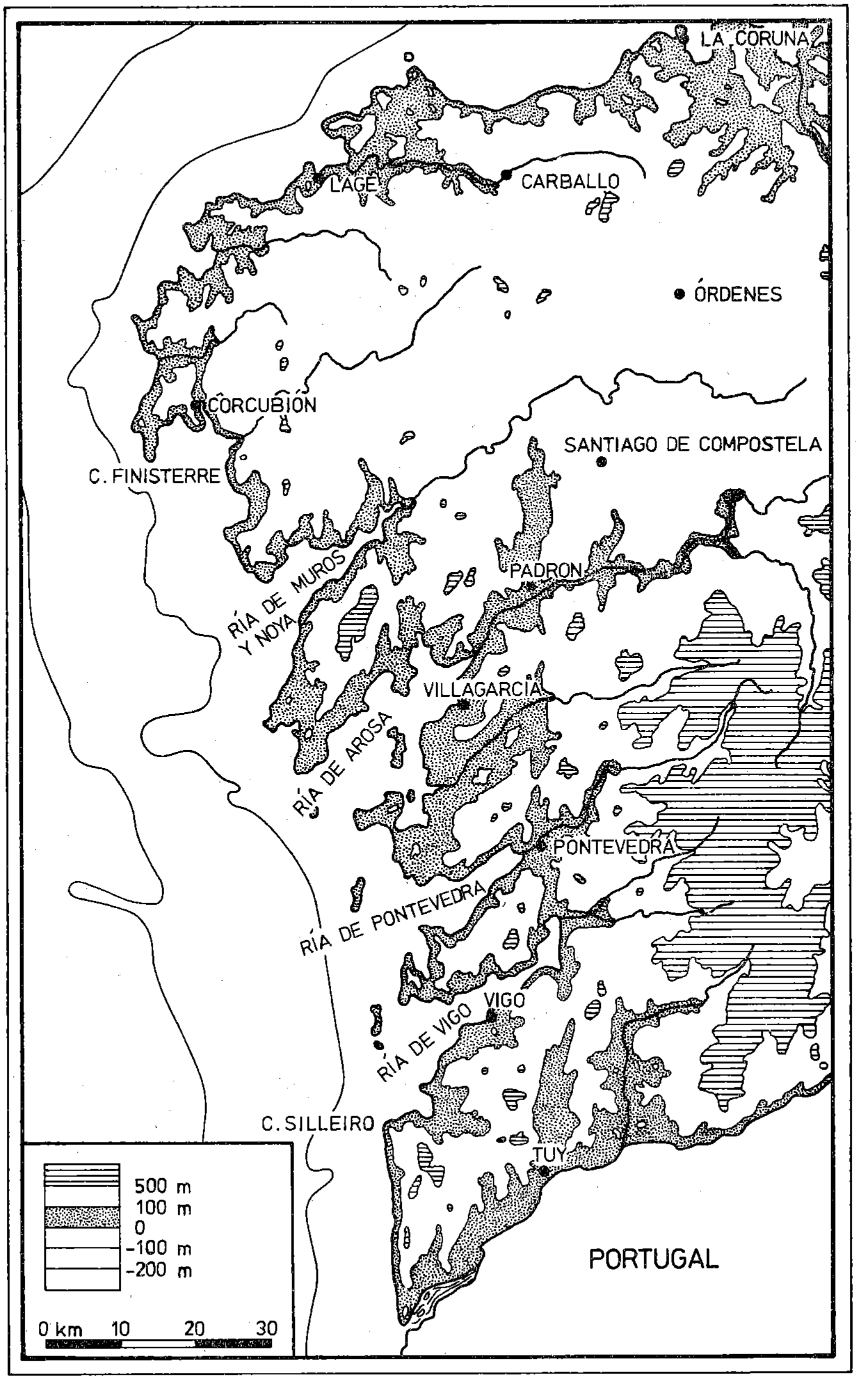


Fig. 1.1 The 'Rías Bajas' of Galicia.

width of the main channel is 3 km. Four larger islands: the islands of Sálvora, Arosa, Cortegada and la Toja, a large number of smaller islands and rocks and several secondary embayments together form a complex topography. The depths are far from uniform: the largest depth recorded was about 70 m, but regions with extensive tidal flats were also found. The total volume of the ria was estimated to be about 4300×10^6 m^3. Further particulars on the bottom topography will be given in a later section (§ 1.5).

From the coastline the country gradually rises towards the low mountains surrounding the ria, up to altitudes of approximately 600 m.

Apart from a number of small rivulets three rivers discharge into the Ría de Arosa. They are, in sequence of importance, the rivers Ulla, Umia and Beluso.

The most important of the numerous towns and villages along the ria is Villagarcia with about 21.000 inhabitants. The most conspicuous human activity on the ria is the cultivation of mussels, by means of rafts from which some 800 ropes are suspended, forming the substrate on which the mussels are growing (see KORRINGA, 1967). No indications of water pollution of any importance were found, although some interference of sewage systems, industry, agriculture or shipping on the waterproperties is probable. But as far as our observations are conclusive, these effects are not important. 'Red water' sometimes occurs in the Rías Bajas, where this phenomenon is called 'purga de mar' (MARGALEF, 1956).

In fig. 1.2 a map is given indicating the names of islands, capes, bays, etc.

1.4 Subdivision of the Ría de Arosa

The topography of the bottom of the Ría de Arosa has been discussed by KOLDIJK (1968). Here we shall mention the points of interest for the study of the water movements. The data are partly from nautical charts, partly from additional observations with a recording echo-sounder.

We may divide the ria into an inner part, a central part, an oceanic part and a southeastern part. The inner part of the ria comprises the water east of the line connecting Punta del Chazo and Isla de Arosa, and north of the sand spit of El Vado. The central part lies between the line Punta del Chazo-Isla de Arosa and the line running from a point halfway between Sta. Eugenia de Ribeira and Palmeira via the islands of Rua, Jidoiro Pedegroso, Jidoiro Arenoso to the southern point of Isla de Arosa. The oceanic part lies to the south of this line and on the east is separated from the southeastern part by a line connecting Isla de Arosa with El Grove. The boundary with the ocean is formed by the rocks and islets north of Isla Sálvora and the line between these islands and Punta San Vicente.

Another subdivision is that between what is called the main channel and the secondary bays. The main channel is the zone of maximum depths stretching from the entrance of the ria between the Isla Sálvora and the peninsula of El Grove in a northern

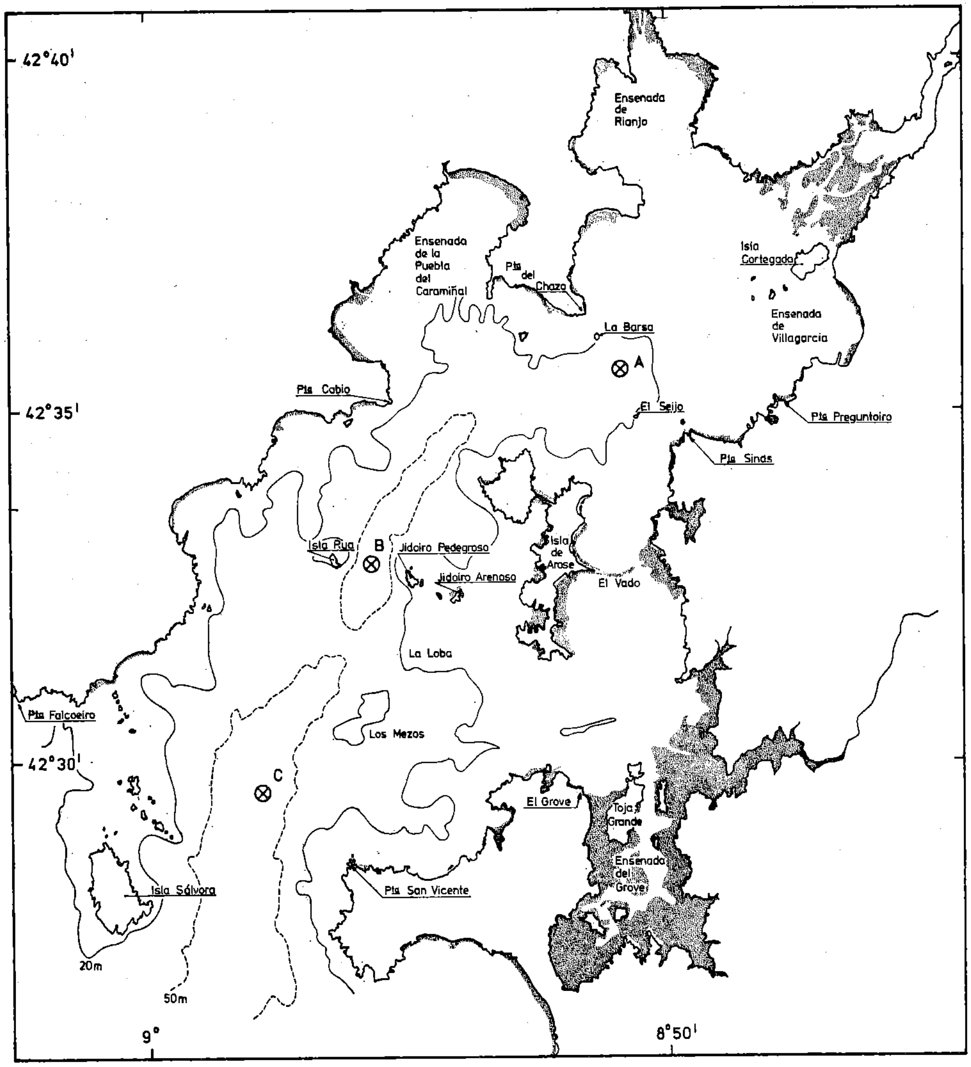


Fig. 1.2 Ría de Arosa.

and northeastern direction towards the estuary of the Rio Ulla. The secondary bays are, along the northwest coast, the Ensenada de la Merced (or Ensenada de la Puebla de Caramiñal) and the Ría de Abanqueiro (or Ensenada de Rianjo) and, along the southeastern coast, the above mentioned southeastern part, consisting of the bay in front of the Rio Umia and the shallow Ensenada del Grove.

In the following table the dimensions are given of the different parts into which the ria can be divided. These data were obtained from the nautical chart, using a planimetre. The figures given are those at mean sea level.

TABLE I. *Dimensions of different parts of the Ría de Arosa*

	surface	volume	average depth
estuary of Rio Ulla	$7.7.10^6$ m^2	26.10^6 m^3	3.3 m
inner part	62.2	685	11
central part	47.0	1118	24
oceanic part	81.0	2351	29
southeastern part	31.8	155	5
whole Ría de Arosa	229.7	4335	19

1.5 Depth distribution

The following data are from the nautical chart no. 924 of the Spanish Hydrographic Service, supplemented with our own echo-soundings. The depths in the nautical chart refer to the mean lowest tide, which is about 2 m below the mean sea-level. The echo-soundings were not reduced to a fixed reference level, as they were made mainly with the purpose of obtaining morphological information on the ria bottom. This means that the depths as recorded by the echo-sounder may be some 3.5 m larger than the corresponding with values of the chart (if observations were made near the time of high water at springs).

The shelf region in front of the Rías Bajas is rather even (c.f. PANNEKOEK 1966). The 70 m isobath lies just in front of the Ría de Arosa. Between the Peninsula of

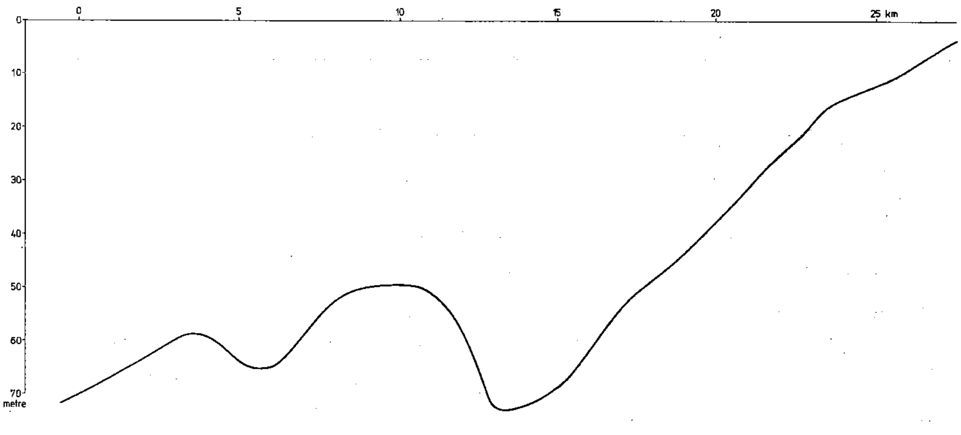


Fig. 1.3 Depth profile along the axis of the Ría de Arosa.

El Grove and the island of Sálvora the main channel joins the ocean. In fig. 1.3 a profile is given of the depths along the axis of the main channel. We see that the depths are not continuously decreasing inwards, but that two 'sills' and two 'deeps' are present. The first sill has been called the 'ocean rise' by KOLDIJK. It is situated between Isla Sálvora and Punta San Vicente. As the sill rises to only some 5 m above the bottom of the deep just behind it, this sill is not an important feature of the topography. The second sill ('central rise') with depths of about 50 m is much more pronounced. This sill is found about halfway between the rocks of Los Mezos and the islet of Rua. Behind this sill a basin with depths of more than 70 m is found which will here be called the 'Rua deep'.

A separate basin lies just north of the village of El Grove. This basin, with depths of about 25 m, is separated from the main channel by a sill connecting El Grove and Isla de Arosa, with depths of about 10 m. In the following this basin will be called the 'Grove deep'.

1.6 Tides

The tide in the Ría de Arosa is of the semi-diurnal type. The amplitudes of the most important tidal constituents at Villagarcia, according to the tidal tables edited by the 'DEUTSCHES HYDROGRAPHISCHES INSTITUT', are shown in the following table:

TABLE II. *Tidal amplitudes in the Ría de Arosa*

diurnal tides:	K_1	amplitude (cm)	9
	O_1		8
	P_1		3
semi-diurnal tides:	M_2		113
	S_2		39
	N_2		22
	K_2		10
	μ_2		3

High and low water occur at Villagarcia only about 5 minutes later than at the entrance of the ria, and the tidal range varies by only some decimetres over its whole length.

From the data given it follows that the tidal range is about 3 m during spring-tide and about 1.5 m during neap tide. At spring-tide high water occurs around 04 and 16 h GMT, at neap-tide around 10 and 22 h GMT.

In fig. 1.4 the tidal curves for spring-tide and neap-tide are given as recorded by the tide-gauge at Punta Preguntoiro. The reference level is that of the nautical chart, within an accuracy of 10 cm. This reference level was established by comparing the tidal observations over a long period with the predicted tides.

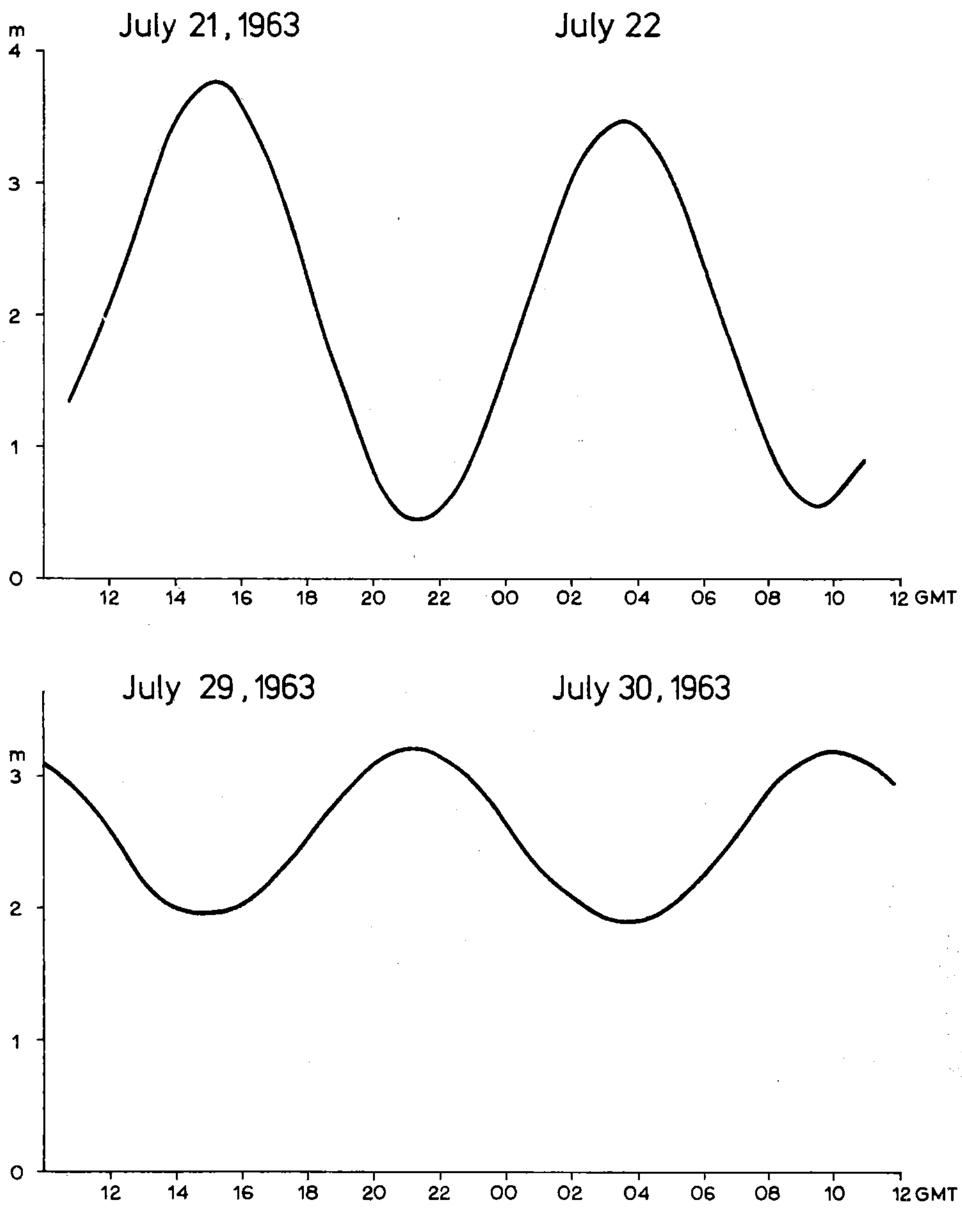


Fig. 1.4 Tidal curves recorded at Punta Preguntoiro.

1.7 Non-tidal sea level variations

Several times small-amplitude, short period oscillations were observed on the registrations of the tide-gauge. Especially around the times of high and low water and during the neap tides these oscillations could be observed. At these occasions it is probably easier to detect such oscillations, owing to the slow variation of sea level. The amplitude of these oscillations was estimated to be not more than some centimetres, the period was between 30 and 45 minutes.

We can only make a guess as to the nature of these oscillations but one might think of seiches, free standing waves of the ria or parts of it. This possibility can be tested by considering the elementary case of a rectangular bay with a uniform cross section with dimensions comparable to those of the part of the ria under consideration. The period of an uni-nodal seiche over the inner part of the ria (regarded as a closed basin) using a mean depth of 11 m and a length of 9.5 km (the distance from the mouth of the Río Beluso to Punta Preguntoiro) is about 30 min. This agrees well with the observed period and therefore seiches are a likely cause for the oscillations.

According to the 'West Coasts of Spain and Portugal Pilot' (HYDROGRAPHIC DEPARTMENT, 1957), westerly winds may cause a rise of the sea-level of 0.4 m and easterly winds may cause a similar lowering. Our own registrations do not permit the determination of such variations, because of instrumental difficulties (see section 5.1). Defects of the tide gauge cannot be ruled out as the cause for observed discrepancies between predicted and observed water levels.

Theoretical investigations of the stationnary wind set-up are presented in section 10.2. From these considerations it can be deduced that a wind of 7 Bft blowing along the axis of the ria would give a maximum set-up over the inner ria of about 8 cm. The contribution of the outer ria would be smaller because of its larger depth. Although nonequilibrium effects (e.g. resonance) and external surges may contribute to a further increase or decrease it is thought that the quoted value of 0.4 m is at least rather uncommon.

1.8 Sea and swell

Information on the occurrence of short-period waves is scarce. In the summer of 1964 the estimated wave height at Punta Preguntoiro was noted every three hours. Furthermore, only casual observations were made.

Table III shows the maximum value of the significant wave height observed at Punta Preguntoiro with different wind conditions. For comparison the values are given for a fully developed sea, according to diagrams presented by GROEN AND DORRESTEIN (1958).

It appears that the observed values for northerly winds do not attain the values of a fully developed sea, and more and more stay below these values with increasing wind

TABLE III. *Some wave data for Punta Preguntoiro*

Wind velocity	Maximum of observed significant wave height northerly winds	Maximum of observed significant wave height southwesterly winds	Significant wave height of fully developed sea
1 m/s	< 0.1 m	< 0.1 m	
2	< 0.1	0.1	0.08 m
3	0.1	0.1	
4	0.2	0.1	0.32
5	0.3	0.1	
6	0.4	—	0.72
7	0.4	—	
8	0.5	—	1.3
9	0.5	—	
10	0.5	—	2.0

velocities. For southwesterly winds there is no appreciable dependence on the wind velocity. It is clear that the small fetch of the wind in the ria limits the wave height that can be attained.

Generally speaking the situation at most other places along the ria is such that the fetch will be not much more than 10 km. Therefore sea waves of more than ½ m high will occur only seldom.

A low swell was regularly observed in the outer ria, but never in the inner ria. CADEÉ (1968) has mentioned an unpublished report of the decrease of swell in the Ría de Arosa made for the port of Villagarcia. According to this study 10 m high swell at the entrance of the Ría (period unknown) will be reduced to 2 m at 13 km from the entrance, that is near the Islet of Rua, and to 1 m at 16 km from the entrance.

The Ría de Arosa is open to the south and southwest, and waves from these directions may enter as swell. Outside the ria waves from western to southern directions occur seldom in summer, but in winter they are observed more frequently, as is shown in the 'Oceanographic Atlas of the N. Atlantic, Section 4' (U.S. NAVAL OCEANOGRAPHIC OFFICE, 1963).

Therefore it is clear that high waves of more than 2 m will mainly be restricted to the outer ria and to the period from December till May.

1.9 Investigations in the Ría de Arosa

The investigations of the oceanography of the Ría de Arosa are part of a larger investigation, covering various aspects of the physical, biological and geological situation. These investigations, carried out during three summer campaigns in 1962, 1963 and 1964 and one winter campaign in 1964 were organized by the Department of Geology and Mineralogy of the State University of Leiden and by the 'Rijksmuseum van Natuurlijke Historie' (State Museum for Natural History) at Leiden, with the cooperation of various other institutes and services in the Netherlands, and with the help of a great number of Spanish authorities and private persons. As regards the physical and chemical oceanography the main contributions came from the Royal Netherlands Meteorological Institute (de Bilt) and the Netherlands Institute for Sea Research (Texel). A full account of the expeditions has been given by BRONGERSMA AND PANNEKOEK (1966).

The results of several investigations made within the framework of these expeditions have already been published: Especially the studies on the geomrophology by A. J. PANNEKOEK (1966b), on the mollusca by G. C. CADEÉ (1968), on the bottom sediments by W. S. KOLDIJK (1968) and on the algal vegetation by M. DONZE (1968) are relevant to the present work.

The observations were made during the following periods:

1962: July 6-August 9
1963: July 1-August 1
1964: January 28-February 7
July 15-August 15

The main camp was established on the small peninsula of Punta Preguntoiro, Villajuan, near Villagarcia. This point offered an excellent base for these observations. In the factory of Don Luis Losada Lago a temporary laboratory was established and the presence of a landing pier facilitated the operation with ships.

Locally hired boats were used, on which the necessary equipment, such as winches, was installed. The following boats were used.

1. the 'Flor da Ponte', a 12 m motor ship, used during all campaigns,
2. motor sloop, used during the second and third summer,
3. occasionally during the third summer a larger salvage steamer was used, mainly for geological work,
4. a rubber boat for work on the rivers and other shallow and sheltered places.

In July 1968 oceanographic observations have been made in the Ría de Arosa by GÓMEZ GALLEGO (1971). Reference is made to these observations where relevant.

1.10 Survey of the present study

Before starting on a description of the oceanographic situation in the Ría de Arosa we shall first consider what might be called the boundary conditions: the meteorology, the hydrology and the oceanography of the bordering ocean waters.

After an introductory chapter on the methods used, the oceanographic situation will be described in the chapters 6 and 7 and, where possible, the observed phenomena and relations will be explained. Special emphasis will be given to the upwelling process that is supposed to play an important part during the summer season. These chapters will be followed by two chapters, in which several physical and chemical properties of the water will briefly be dealt with.

Finally a conclusion will be given in two chapters, the former giving a theoretical model, approximating the circulation and mixing processes, the latter relating the present investigation to other data.

2. Notes on the meteorology of the Ría de Arosa and surroundings

2.1 Introduction

It is not the intention to give in this chapter a complete climatological description of this area, but rather some information that is thought to be necessary for a good interpretation of the results of the various investigations carried out in this area. As the major part of these investigations were made during a number of summer campaigns, special attention was given to the situation during the summer months. For the greater part this study consists of a compilation of data from various sources. In the succeeding paragraphs reference will be made to these publications. Some information may also be found in publications of the U.K. HYDROGRAPHIC DEPARTMENT (1957) and the U.K. METEOROLOGICAL OFFICE (1941) and the Spanish SERVICIO METEOROLÓGICO NACIONAL (1952, 1968). These data are supplemented with meteorological observations made at the site of the base camp at Punta Preguntoiro during the investigations. Also occasional observations, made on and around the ria, have been used to get a picture of the weather in this area.

2.2 Observations at Punta Preguntoiro

The small peninsula of Punta Pregnuntoiro protrudes in the inner part of the ria in a NNW direction from the shore, which here follows a NE-SW direction. On this peninsula is a building which was enlarged during the course of the years and which in 1964 covered about one third of the whole area. Although the place where the observations were made was carefully selected, so that the influence of this building was judged as small as possible, and although also the distance from the shore was chosen as large as the situation permitted in order to minimize the effect of coastal disturbances of the windflow, we realize that the site was not an ideal one for meteorological observations. This is one of the reasons why our observations can only give a rough picture of the weather situation.

The observations were initially meant to give additional information for the oceanographic observations on the ria and only later on they appeared useful for a short description of the weather. As these observations were some of the many tasks to be done by a varied group of people, they were planned to be made during daytime only. On all days except Sundays observations were made at regular hours, namely at 08, 11, 14, 17 and 20 h GMT (09, 12, 15, 18 and 21 h offical Spanish time).

2.3 Instruments and methods

The observations consisted of measurements of air temperature and air humidity with a hand wet-and dry-bulb thermometer set, wind measurement with a hand-anemometer at about 2 m above ground-level and some 4-7 m above sea-level and an estimate of the cloud amount, with additional notes on the types of the clouds and other relevant details of the weather situation such as the presence of fog, rain or drizzle. In 1963 and 1964 a rain-gauge was used as well. The collected rain was measured once a day, or more often if necessary.

Once, on July 12, 1963 during a heavy rainstorm, the gauge overflowed. As the time at which this happened and the duration of the rain storm are known approximately, the total precipitation could be estimated from the observed intensity of the rain.

On board of the ship the wind force was estimated in the Beaufort-scale, but in 1964 a hand-anemometer was used as well.

2.4 Annual variation of the meteorological elements

In annex I a number of tables and graphs is brought together, showing the annual variation of wind, temperature, precipitation and humidity in the coastal area of northwestern Spain and over the adjoining sea. An important part is played by the seasonal variation in the pressure distribution. During summer the influence of the Azores high usually extends to this region and as a consequence depressions are infrequent and northerly winds predominate, while the precipitation reaches a minimum. During winter, when the Azores high has a more southern position, the passage of depressions and fronts occurs more frequently, giving a more variable meteorological situation.

The predominant northerly winds in summer, the so-called 'Portuguese trades' extend along the whole western coast of the Iberian peninsula. Close to the shore orographic influences play a part. The frequency of gales is highest in winter and southerly and southwesterly gales are then most frequent. As may be expected the annual variation of temperature at sea is much smaller than on land. On land differences in height also have their influence, as can be seen from the comparison of the figures for Pontevedra and Santiago de Compostela, the latter being 1 or 2 °C lower throughout the year. Along the coast frost as a rule only occurs during December, January and February as is shown by the Pontevedra data.

The annual variation of precipitation is conspicuous. The total annual rainfall in this area is the highest of the Iberian peninsula. Of course considerable variations in individual months and years are possible. At sea the same tendency of the annual variation was found, although this cannot be expressed in quantities of precipitation, but in frequency of precipitation. As regards the variation caused by differences in height above sea level we may refer to LAUTENSACH (1951).

The relative humidity, as represented by the data from Pontevedra, shows higher values in winter than in summer. This variation is a consequence of the variation in air temperature. The specific humidity, on the contrary, has a maximum in summer.

Fog at sea shows a maximum in August-September. To a certain extent this is also reflected in the data of Pontevedra. This high frequency of fog (and also of haze) during the summer months is caused by the relatively low temperatures of the sea water along the coast which are brought about by upwelling of water from greater depths.

In agreement with the proximity of the high pressure area and the low precipitation the cloudiness is minimal in August, at sea as well as on land.

2.5 Diurnal variation of the meteorological elements during the summer

The greater part of the investigation of the Ría de Arosa was carried out during summer. For a better understanding and interpretation of the results, and especially as a background for the study of the oceanographic situation in the Ría de Arosa, it is useful at some length to go into the diurnal variation of the meteorological elements in summer. In this connection the variations in the area of the ria itself are the most important, but they have to be supplemented by data from stations over a larger area. During the different summer campaigns meteorological observations were made at the site of the camp at Punta Preguntoiro. No regular observations were made during the night.

However, these data, if compared with data from official meteorological stations still give a useful picture of the diurnal variations. The observations made at Punta Preguntoiro during the summer campaigns were combined and diurnal frequency distributions of wind, temperature, humidity and cloudiness were calculated. They are presented in annex II.

In the beginning of the morning northerly winds are predominant over the Ría de Arosa, often with wind forces of 4 Bft and more, whereas there is also a high percentage of calms. In the course of the day the frequency of northerly winds decreases and the percentage of calms drops. On the other hand the frequency of southwesterly and westerly winds increases during the course of the day. In the evening the frequencies of northerly winds and calms increases again. The onset of the northerly winds often happens quite suddenly. Often, within an hour a change from a light southwesterly wind to a fresh or even strong northerly breeze was observed. Such a diurnal variation of the wind is very marked along the whole Galician coastline. The main wind directions, however, are different for each station. This variation, which is clearly a case of alternating land and seabreezes is perhaps also influenced by orographic effects.

As no observations were made during the night, the daily temperature range cannot be given for Punta Preguntoiro. From data published by the METEOROLOGICAL OFFICE (1944) and by SEMMELHACK (1932) we see that there is a decrease of this range when going seaward, because of the moderating effect of the sea water. It may be assumed that the temperature range at Punta Preguntoiro will be somewhere between the values of Pontevedra, lying at the head of the Ría de Pontevedra and of Vigo, which lies more exposed, about halfway the Ría de Vigo. These ranges are 13.3°C and 7.8 °C in July and 13.3 °C and 8.1 °C in August respectively. So the mean minimum temperature in summer at Punta Preguntoiro is estimated to be about 13 °C, that is about 10 °C below the observed maximum of 23 °C. The diurnal variation depends on the weather situation; as can be seen from the data presented in the annex II the variation is larger with northerly winds. The relative humidity shows a minimum around mid-day, which is on the average lower for northerly winds than for southwesterly winds. Both these effects of wind on temperature and humidity illustrate the importance of continental and maritime influences. The variations of the relative humidity are determined not so much by variations in the water content of the atmosphere as by the diurnal variation of the temperature. The vapour pressure shows a maximum at mid-day (17.5 mb) and low values in the morning and evening (16.5 mb). The observations show the occurrence of fog to be most frequent during the morning hours, but the number of observations is too small to give a mean frequency. However, data from La Coruña and Viana do Castelo (N. Portugal), given by the METEOROLOGICAL OFFICE (1944) can be taken as representative. They show that a visibility less than $\frac{1}{2}$ nautical mile occurs in summer with a frequency of 13 % and 26 % at 07 h GMT, and of 4 % and 5 % at 13 h GMT respectively. The average cloudiness only shows a slight diurnal variation. If the conditions with northerly and southwesterly winds are taken separately, we find higher cloudiness for the latter case, while also the diurnal variation is more pronounced under these conditions.

2.6 Local differences of the weather over the Ría de Arosa

The meteorological data for Punta Preguntoiro, as given in the previous sections, cannot directly be applied to the situation in other parts of the Ría de Arosa. It very soon appeared that considerable differences may occur over the ria. These may be due to differences in the orographic situation and to various influences exerted by the land and the water.

As it was impossible to investigate these local differences in detail, we can only study them from more or less occasional meteorological observations made aboard the ships during oceanographic work and by groups that for some time were working at certain other points of the ria. These observations were worked up as follows:

Assuming that in all cases with winds at Punta Preguntoiro from northern or southwestern directions the general pattern of the meteorological elements is always more or less the same, we may combine all observations from the different parts of the ria, for both wind situations separately, in a regional pattern in order to get a, be it more or less qualitative, picture of the local differences for both cases. In order to eliminate the diurnal and interdiurnal variations as far as possible, the differences between the observations and the simultaneous observations at Punta Preguntoiro were taken and then the means of these differences were computed for the various parts of the ria. For temperatures and relative humidities the differences with properly interpolated values at Punta Preguntoiro were determined, if observations had not been made at standard hours. The observed wind was compared with the wind at Punta Preguntoiro at the nearest observation hour.

2.6.1 Conditions during northerly winds at Punta Preguntoiro

All cases with northerly winds have been combined in the way described in the preceding section. Exact numerical data are not given here, as the number of observations is not sufficient to give more than a rough picture. The results have been used, however, to estimate regional differences of the evaporation (see section 3.5). The tendency that appears in the variation of temperature, humidity and wind will be described below in a qualitative way.

It appears from these data that there is a decrease of wind force when we go from the northern to the southern part, and from the eastern to the western part of the ria, while at the same time a change from northeastern to more north western directions is found. Apart from effects of the orography on the land winds (channelling of the wind in the valley of the Rio Ulla) and of differences in the land-sea distribution on the land winds the stabilizing effect of the large, rather cold water surface can be regarded as one of the reasons for these differences of the wind. The higher stability brings about a lower velocity and a simultaneous backing of the wind near the surface.

A tendency of decreasing air temperatures from northeast to southwest is manifest, with exception of the shallow, enclosed region in the southeast. No doubt this general trend of decreasing temperatures is caused by the cooling of the air by the water surface. The air temperatures at the entrance of the ria were found to be, on an average, 1° to 1.5° below the simultaneous temperatures at the reference station at Punta Preguntoiro. The area enclosed between the Isla de Arosa, El Grove and the mouth of the Ria Umia had temperatures which, on an average, were about 1° higher than those at the reference station. The relative humidity in the northwestern and the southwestern part of the ria was high, as compared with Punta Preguntoiro, while in the area near El Grove the relative humidity is lower. The cooling of the air over the sea gives lower saturation values and thus the relative humidity increases, even if no evaporation takes place at the sea surface. The higher temperatures and lower relative humidities of the southwestern parts of the ria as compared with Punta Preguntoiro indicates that the influence of the low water temperatures is already apparent at the latter station.

2.6.2 Conditions during southerly and westerly winds at Punta Preguntoiro

Similar to the northerly winds, the southerly and westerly winds appear to be strongest in the northeastern parts of the ria. In the southern part the westerly winds also appear to have a predominant northerly component, while they have a southerly component in the northern parts. As far as these winds can be regarded as sea breezes we may expect the irregularities in the distribution of land and sea, and also the presence of mountains, valleys and plains to influence the direction and speed of the local wind. The cold water of the ria brings about a subsidence of the air, which agrees with the more or less divergent motions of the wind at the surface.

As far as the winds are caused by larger scale disturbances, e.g. the presence of a low-pressure area north of the ria, the above-mentioned influences may be of less significance. But differences in friction over land and sea will result in a smaller deflection of the wind direction in the south, i.e. a wind direction more parallel to the isobars, where an almost uninterrupted path is open over the sea. However, as this lower friction would also give higher wind speeds in the south, which is in contrast to our observations, this latter reasoning fails to explain all aspects of the observed differences.

The temperatures of the air over almost the whole ria on an average are between 2° and 3° lower than those at Punta Preguntoiro. Only the eastern shore has temperatures that are more or less equal to those at the reference station. The explanation is that with southerly and westerly winds the air is advected from the sea and that its properties are changed only when it flows over land. This is confirmed by the pattern of relative humidity, with low values along the eastern shore.

2.7 Weather during the different campaigns

Data on the weather during the different months can be found in the regular Spanish meteorological bulletin (SERVICIO METEOROLOGICO NACIONAL, var. years). Among other data, maps are given of monthly temperature and precipitation anomalies. These maps give a general impression of the deviation of the weather during that particular month from the average over a large number of years. This leads to the following characterization of the summer periods during which the investigations have been carried out:

1962, July:	temperature approx. normal, precip. normal
August:	temperature approx. normal, precip. slightly below normal
1963, July:	temperature approx. normal, precip. normal
1964, July:	temperature approx. normal, precip. slightly below normal
August:	temperature approx. normal, precip. slightly below normal

There appears to be no reason to suppose that the weather during the three successive summer campaigns would not be representative for climatologically normal conditions. As regards the weather during the winter campaign from January 28 till February 7, 1964 it can be said that in both January and February the temperature was somewhat above normal, whereas the precipitation during January was far below the normal value. Although in February the monthly precipitation was more than the normal average, this had no influence on the conditions encountered during the investigations.

In the following paragraphs some more details about the weather during the different observation periods will be presented, in order to give a general idea of the circumstances under which the various investigations have been carried out.

2.7.1 Weather during the period July 9 till August 4, 1962

The weather during this period was characterized by the occasional passage of cold fronts until July 27, when the Azores high started to extend its influence towards the Iberian peninsula. Another passage of a cold front occurred at August 2.

At the end of July maximum values of the temperature were reached. The highest value observed at Punta Preguntoiro was 30.2 °C on July 31. Rain and drizzle were observed on 5 days during this period, viz. on July 9, 13, 18 and 20, and on August 1. No measurements of the precipitation were carried out, but the highest amount measured in 24 hours at the neighbouring station Zamar de Rubianes was 6.0 mm on July 10.

During the first part of the period, until July 26, winds from southwestern directions predominated afterwards northerly winds were more frequent.

2.7.2 Weather during the period July 1 till 30, 1963

Apart from some influence of occasional local low pressure areas, the weather was characterized by a more or less continuous period of undisturbed weather, until July 12, when a depression developed on the peninsula, which was accompanied by a short interruption of the generally fine weather. A new period of fair weather was brought to an end by the passage of a cold front on July 25. After this the weather showed a greater instability.

On the 20th the highest temperature, viz. 28.0 °C, was observed at Punta Preguntoiro. Precipitation was observed on 6 days (July 9, 10, 11, 12, 13, 25), with a maximum of about 19 mm (estimated) on July 12. The winds generally showed the pattern of diurnal variation as discussed in paragraph 2.5. Strong winds were observed on July 19 (more than 12 m/sec).

2.7.3 Weather during the period July 14 till August 14, 1964

During the first ten days, due to the influence of high pressure areas, the weather was generally fine. From July 24 the atmosphere showed a greater instability, while on July 27th a cold front passed from the north to the south; but afterwards the high pressure area regained its influence. On August 7 the passage of another cold front

was the beginning of a period of less constant weather, with occasionally showers.
A highest temperature of 31.2 °C was observed at Punta Preguntoiro on August 4. Precipitation was observed on 4 days (July 25, August 12, 13 and 14), with a maximum of 3.8 mm on July 25. Between July 27 and August 3 winds remained northerly almost permanently. For the rest of the period there was a more or less regular diurnal alternation of northerly and southwesterly winds.

2.7.4 Weather during the period January 28 till February 7, 1964

For the major part of the time the weather was influenced by high pressure areas; only on January 29 cloud systems penetrated this area. As a result the weather was generally warm, with less precipitation than is normal for this time of the year.

The maximum temperature observed at Punta Preguntoiro was 16.9 °C on February 3. Other stations around the Ría de Arosa observed still higher temperatures (Herbón Padrón 21.0 °C on February 3). The observations made at Punta Preguntoiro during this whole period give the following average daytime temperatures:

08 h: 8 °C; 11 h: 10 °C; 14 h: 14 °C; 17 h: 14 °C.

In January the amounts of the total precipitation at the stations around the ria were between 24.9 mm (Zamar de Rubianes) and 8.0 mm (Sta. Eugenia de Riveira), the major part of this precipitation having fallen in the first part of the month. The values of February were much higher, but of this precipitation all fell in the second half of this month. The winds blew predominantly from northeastern directions, with velocities up to 10 m/sec. Apart from January 29 and 30 there was a clear sky nearly all the time.

3. Hydrology of the drainage basin of the Ría de Arosa

3.1 Introduction

The oceanographic conditions of the Ría de Arosa can only then be properly understood if the fresh water run-off, especially the discharge of the two larger rivers, the Ulla and the Umia, is known. The large annual variation of the precipitation, for which figures have been given in annex 1, of course has a considerable importance for the regime of the rivers. The large annual variation of the evaporation also has to be considered. As a consequence large seasonal differences occur in the run-off.

The hydrology of Galicia has been discussed by NONN (1966) and the figures quoted by this author are of special importance to our study.

3.2 Water balance

In a publication by C. W. THORNTHWAITE ASS. (1964) on the average climatic water balance of Europe data have been given for Santiago de Compostela, Pontevedra and Lugo. The procedure followed in this publication for the calculation of the different elements of the balance has been given in an earlier study (THORNTHWAITE AND

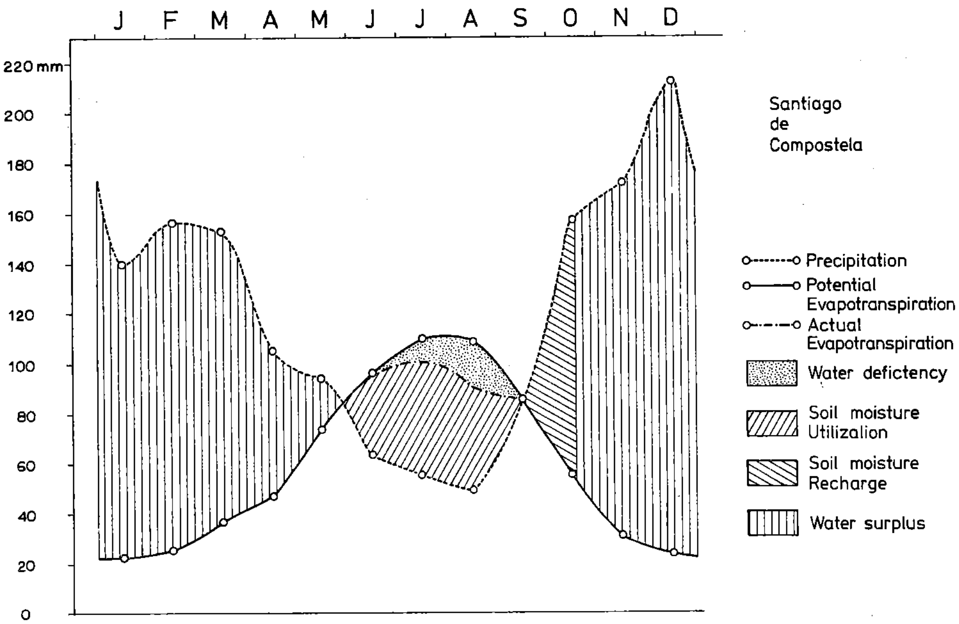


Fig. 3.1 Annual variation of the elements of the water balance at Santiago de Compostela. (Thornthwaite Ass. 1964).

MATHER, 1957). The reader is referred to this latter publication for the details of the method. Here we may note that only data were used on temperature and precipitation and information on the water holding capacity of the soil, that is the 'mass of water retained by a previously saturated soil when free drainage has ceased'. The main points are that the potential evapotranspiration is calculated from temperature data and data on the duration of the sunlight for the particular month and latitude. The water holding capacity of the soil had to be estimated (in the present case 300 mm), whereas actual evapotranspiration was deduced from the potential evapotranspiration by application of a reduction factor that is equal to the actual soil moisture to the water holding capacity ratio.

It is apparent that this method gives only a rather general estimate. In his review of the climate of the Iberian peninsula LINÉS ESCARDÓ (1970) quotes an evaluation by Elias Castillo of the different methods for computing the potential evatranspiration. According to this study the Penman figures are up to 20 % higher than those of Thornthwaite, while the former method appears to give more correct results in the Mediterranean area. So there is every reason to look at the data of fig. 3.1 only as indications of the general trend. It should furthermore be noted that the precipitation figures are averages over the period 1901-1930, whereas those of annex I refer to 1861-1900.

A point which deserves special attention is that, although the precipitation is increasing considerably in September and October, there still is no water surplus, as the soil moisture storage is not yet at field capacity. Therefore the run-off can be expected to lag behind the precipitation.

The ratio of annual water surplus to annual precipitation is found to be 0.524, 0.529 and 0.488 for Santiago, Pontevedra and Lugo respectively. This is in good agreement with the data presented by URIARTE HUMARA (1966), who for the whole northern and northwestern coastal zones of Spain gives an average annual run-off which is 0.516 times the average annual precipitation.

3.3 General notes on the regime of the Galician rivers

No data on the run-off of the rivers Ulla and Umia could be found. However, some figures on other Galician rivers have been presented by NONN (1966). Especially the data on the rivers Lerez and Tambre are of interest, as the drainage basins of these two rivers are situated to the south and north, respectively, of those of the Ulla and the Umia. The Rio Lerez is the most important tributary to the Ría de Pontevedra, and the Rio Tambre is the main river flowing into the Ría de Muros y Noya. As furthermore the dimensions of the catchment areas of these two rivers are not much different from those of the Ulla and Umia, we may suppose that the regime of the latter two rivers may be estimated from the data on the Lerez and Tambre.

The following figures have been given by Nonn for the discharge of the Rio Tambre (catchment area 1100 km^2):

'Normal' year	23.7 m^3/s
'Wet' year	41.4
'Dry' year	15.8

It was estimated that under normal conditions the rivers of western Galicia have a discharge per unit surface area of the catchment basin of 21-23 litre/s km^2. For smaller rivers, however, this figure is perhaps somewhat larger, as is indicated by the value for the Rio Oitabén, discharging into the Ría de Vigo, with a catchment area of 333 km^2. For this river a value of about 41 litre/s. km^2 is found from the data given by SAIZ ET AL. (1957).

Nonn gives the quotient of the monthly mean discharge (per unit of time) and the annual mean discharge as a monthly discharge coefficient. In table IV his discharge coefficients for the Tambre and the Lerez are given together with the average values of these two coefficients.

TABLE IV. *Annual variation of the discharge coefficients of some Galician rivers according to* NONN (1966)

	Tambre	Lerez	Average
Jan.	2.18	1.65	1.92
Febr.	2.43	2.16	2.30
March	1.88	2.09	1.98
April	1.15	1.2	1.18
May	0,715	0.73	0.72
June	0.22	0.455	0.34
July	0.19	0.19	0.19
August	0.163	0.156	0.16
Sept.	0.15	0.198	0.17
Oct.	0.36	0.39	0.38
Nov.	0.593	0.64	0.62
Dec.	1.71	1.92	1.82

In October and November the discharge is still below the annual mean, although the precipitation already attains high values.

According to Nonn the mean annual run-off of the Tambre and the Lerez is 62.5 % and 60 % respectively of the mean annual precipitation. This is somewhat more than the figure of 52 % given in the previous paragraph for the ratio of water surplus to precipitation. Although different reasons might be advanced why a close agreement between these different figures cannot be expected, it might be advisable not to attribute an accuracy better than 20 % to the values for the river discharge given in this section.

3.4 Measurement of the discharge of the Rio Ulla and Rio Umia

On three occasions the discharge of the Rio Ulla was measured, twice in summer (on July 11, 1963 and on July 22, 1964) and once in winter (on February 8, 1964); the discharge of the Umia was measured once (on August 12, 1964).

In principle the usual procedure was that a cross-section was chosen where the river flow was thought to be relatively undisturbed. The area of this cross-section was determined by measuring the water depth at regular intervals along the section from one bank to the other. Surface velocities of the water were measured by means of wooden floats. These measurements, made at various distances from the river banks, showed the lateral variations of the current.

The vertically averaged current was estimated by assuming a current profile that can be represented by the formula

$$v_z = v_h \left(\frac{z}{h}\right)^{\alpha}$$

where v_z is the current at a distance z above the bottom and v_h is the current at the surface, where $z = h$. The exponent α has to be determined empirically. In the literature for a number of rivers values of α between 1/5 and 1/7 have been given (VAN VEEN, 1937). DRONKERS (1964) compared the empirical relation with theoretical relations based on the logarithmic velocity profile. For a section of the river Rhine good agreement was found between both relations.

Here we shall assume a value for α of 1/6 which for the vertically averaged current velocity gives:

$$\bar{v} = \tfrac{6}{7}\, v_h$$

By the above method the following results were obtained:

river	location	date	discharge
Ulla	Puentecesures	June, 11, 1963	11.1 m^3/s
Ulla	Puentecesures	February, 8, 1964	17.7 m^3/s
Umia	Barrantes	August 12, 1964	2.0 m^3/s

On July 22, 1964 discharge measurements were made for the Rio Ulla near Herbón. Here the water flows over a weir. By measuring the water level at the weir, as well as the width of the overflow, the discharge was estimated to be 9 m^3/s, using the formula

$$Q = 0.38\, Bh\sqrt{2gh}$$

where Q is the discharge, B and h the width and the depth of the water at the overflow, and g the gravitational acceleration. This formula is a simplified form of a general formula for stream gauging weirs

$$Q = c_D(\tfrac{2}{3})^{3/2} \sqrt{g}\, BH^{3/2}$$

where c_D is a discharge coefficient and H is the total head of the flow (WORLD METEOROLOGICAL ORGANIZATION 1971). The coefficient c_D depends on the shape and roughness of the weir's crest and is usually close to unity. H is somewhat greater than h, because of contraction of the water flow close to the weir.

So if h is used instead of H we should have to apply a correction factor. However, in view of the rather crude measurements of the values of B and h and of the irregular shape of the weir's crest an accurate determination of Q is not possible, and therefore the simplified formula was used in which H is replaced by h and $c_D = 1$.

The values given for the discharge of the Ulla are measured upstream of the mouth of the major tributary river, the Rio Sar. For the total discharge of the Ulla the results therefore have to be multiplied by a factor which is approximated by the quotient of the total area of the catchment basin of the Ulla and that of the catchment basin upstream of the measuring point. This factor was found to be 1.15. For the discharge of the Umia such a factor was not applied, as no important tributaries flow into this river downstream of the measuring point.

The final results are:

Rio Ulla	July 11, 1963	discharge	12.8 m^3/s
	July 22, 1964		10.3 m^3/s
	February 8, 1964		20.4 m^3/s
Rio Umia	August 12, 1964		2.0 m^3/s

The summers of 1963 and 1964 may be regarded as normal summers from the meteorological point of view. Therefore an estimate of the average discharge of the Ulla for July is $11\frac{1}{2}$ m^3/s. If we apply the figures for the monthly variation of the discharge given in section 3.3, the normal annual discharge is estimated to be 61 m^3/s, and the normal discharge in February to be 140 m^3/s. Comparison of this latter figure with the value measured in February 1964 shows that at that moment the conditions were not representative for a normal winter situation. This was also mentioned in the discussion of the meteorological conditions.

To check our results we may calculate the discharge per unit surface area of the catchment basin. As mentioned in section 3.3 a value of 21 to 23 litre/s km^2 is considered normal. The catchment areas of the Ulla and the Umia have been estimated to be 2.780 and 430 km^2, respectively, which gives values of about 22 and 26 litre/s km^2, respectively. These results appear to be satisfactory.

3.5 Evaporation

The potential evapotranspiration in the drainage basin of the Ría de Arosa is high, as is shown in the example of Santiago de Compostela where in July it attains a value of about 110 mm (according to the Thornthwaite method of computation), that is an average of 3.5 mm/day. So in the hydrologic balance of the Ría de Arosa the evaporation may be important, especially if we consider the rather low values of the river discharge.

Evaporation can be estimated from meteorological data. Although this can only give rather coarse figures, it was thought to be better than to take available figures from inland stations such as Santiago de Compostela, because of differences in altitude and maritime influences. ANADÓN ET AL. (1961) considered even the data of the Vigo observatory to be less applicable for the determination of the evaporation over the Ría de Vigo.

Usually evaporation is estimated from the difference between humidity over water and in the air at some higher level, using a transfer coefficient for vapour, $\boldsymbol{D}_w$ (SELLERS, 1967). This relation is

$$E = \rho_{\text{air}} \boldsymbol{D}_w (q_s - q_z) = \rho_{\text{air}} \boldsymbol{D}_w \frac{0{\cdot}622}{p} (e_s - e_z)$$

where ρ_{air} is the density of the air, q is the specific humidity at the sea surface s and at observation level z and p is the total pressure, which at the sea surface can be taken as constant. A theoretical consideration learns that close to the sea surface $\boldsymbol{D}_w$ may be expected to be a linear function of the wind speed W and that it is furthermore determined by the surface roughness. So

$$E = \mathrm{f}(W)(e_s - e_z) = (c_1 + c_2 W)(e_s - e_z)$$

The coefficients c_1 and c_2 depend on the height for which the wind and vapour pressure of the air have been determined, the stability or instability of the air, the fetch of the wind and on the situation of the measuring station with respect to the coastlines and the wind direction. A review of the different problems involved has been in a report of the World Meteorological Organization (GANGOPADHYAHYA ET AL. 1966).

Here we shall use a relation based on the work of ROHWER (1931) and adapted by LAEVASTU (1960), so that it should give the evaporation with wind velocities deduced from Beaufort estimates, which are valid for about 8 m above sea level and temperature and humidity observations aboard the ship. The relation is

$$E = (0.26 + 0.077\,W)(0.98\,e_s - e_z)$$

where E is in mm/day, W in m/s and e_s and e_z in mbar. The measurements at Punta Preguntoiro were made at 4-7 m above sea level (depending on the tide). The difference this makes in the estimate is considered to be small compared with other inaccuracies.

In this formula we see the use of 0.98 e_s instead of e_s for the vapour pressure directly above the sea surface. This factor takes into account the effect of near-surface cooling.

The use of the above-mentioned Rohwer formula implies the introduction of a transfer coefficient $\boldsymbol{D}_w$, which depends on the wind speed according to

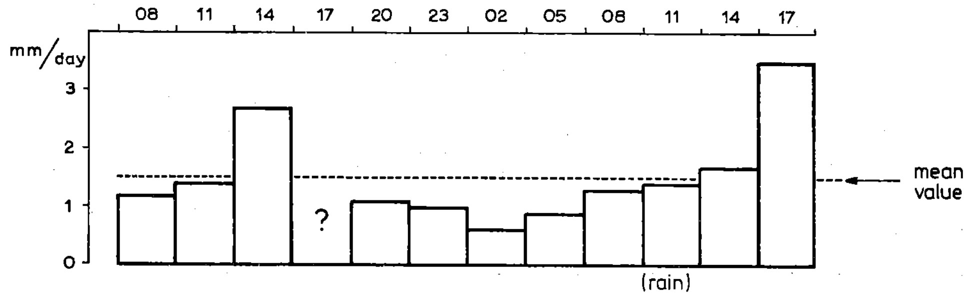
$$\boldsymbol{D}_w = (\boldsymbol{a} + \boldsymbol{b}W) = 0.38 + 0.114\,W$$

for $\boldsymbol{D}_w$ in cm/s and W in m/s

For a level of 2 m Sellers gives values for $\boldsymbol{a}$ ranging from 0.45 for a certain type evaporation pan to 0.07 for a large lake, while $\boldsymbol{b}$ is nearly constant and has a value of about 0.2. For measurements representative of a higher level we may expect a somewhat lower value of $\boldsymbol{b}$, as the wind speed increases upwards. The reason for a relatively high value of $\boldsymbol{a}$ appears to be the larger scale turbulent and convective motion that may be expected to contribute to $\boldsymbol{D}_w$, as calculated over a thicker layer.

Another point to be considered is the question which values of W, e_s and e_z have to be used in the formula. The formula used here has been adapted empirically to give an average daily evaporation rate from diurnal mean values. However, it is difficult to see what will be the effect of the different ways in which the diurnal variation of W, e_s and e_z may be correlated. Furthermore meteorological observations were made only between 08 and 20 h GMT, which means that diurnal mean values of the meteorological parameters cannot be calculated but can only be estimated. Only at two occasions meteorological observations were made during a complete 24 hour period. The first of these series, at Punta Preguntoiro on July 11-12, 1963, was made during a period of winds and the second series was made near La Toja during a period of N winds. From these series the rate of evaporation at 3-hourly intervals was estimated with the above-mentioned formula. Furthermore in the latter case the estimated mean water temperature of the whole bay of El Grove was used for calculating the value of e_s instead of the water temperature right at the measuring station, as here advective variations of water temperature cause an extra diurnal variation.

The diurnal variation of the evaporation rate is given in fig. 3.2. It may be seen that the evaporation in the second case, because of the higher wind velocities and of the lower humidities, is much higher than in the first case. In the first case the evaporation during the period 20-08 h GMT is below the 24 hour mean value, as might be expected because of the higher relative humidity of the air during the night. The evaporation



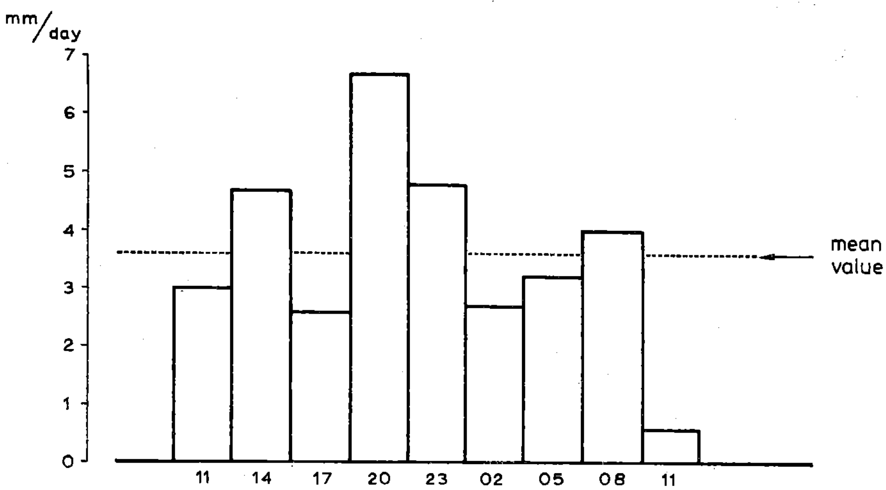


Fig. 3.2 Diurnal variation of the evaporation calculated at two occasions for points in the Ría de Arosa.

calculated from the 08 till 20 h meteorological data is 1.8 mm/day, which is 0.3 mm/day higher than the daily mean. In the second case the situation is complicated by the effect of strong winds during the night. Here the evaporation estimated from the 08 till 20 h meteorological data is the same as the whole day mean (3.6 mm/day).

In the situation of alternating land and sea winds the evaporation will be depressed by the sea wind during the day, and will be increased by the land wind during the night. Therefore no rule can be given for the diurnal variation of the evaporation rate. We shall use all the data from 08 till 20 h GMT in our estimate, although we realize that, judging from both series investigated here, the results may have an error which varies according to the situation, but which may be of the order of 20 % or 0.3 mm/

day (whatever the largest). For the computation of the mean rate of evaporation we may either determine the evaporation of each day and take the mean of all these values or the mean rate from the mean values of W, e_s and e_z. As the wind velocity and the humidity will be correlated to some extent, there will be a difference between these results. However, comparison of both methods for short periods did not give differences more than 0.1 mm/day. This result is encouraging, as for estimating the evaporation over the whole ria we cannot use the first method, because the necessary meteorological data (regularly obtained data at different points of the ria) are lacking, so only the second method (using mean values and the findings of section 2.6) can be employed. Therefore we shall make our estimates from the overall mean values. The accuracy of this result will be sufficient for the purpose of this investigation.

For Punta Preguntoiro we found

$\bar{e}_s = 20.1$ mbar
$\bar{e}_z = 17.2$ mbar
$\overline{W} = 4.7$ m/s

And this gives a mean evaporation $E = 1.6$ mm/day.

This is a rather low value, which may still be too high because of the omission of the night data. Apparently the influence of comparatively low water temperatures is responsible for this result.

The evaporation at Punta Preguntoiro is not representative for the evaporation over the rest of the ria. As seen in section 2.6 the velocity of the wind and the humidity of the air change significantly over the length of the ria. The same is the case with the surface temperature of the water. In order to find estimates for the evaporation in different parts of the ria the segments shown in section 10.6 (fig. 10.3) will be considered separately. The values of the meteorological parameters are only very rough estimates, based on a small number of measurements that were made simultaneously with those at Punta Preguntoiro. From the average differences with the values observed at the latter station the values of $\overline{W}$ and $\bar{e}_z$ were estimated, while those of $\bar{e}_s$ were taken from the temperature distribution shown in fig. 6.2.

It appears that $\bar{e}_s$ and $\bar{e}_z$ are very close together. In that case it is better to use $\bar{e}_s$ instead of 0.98 $\bar{e}_s$ in Rohwer's formula (the factor 0.98 giving a correction for the near surface cooling).

Thus we obtained the following results (Table V):

It is doubtful whether these differences between the segments are realistic. However, the evaporation apparently is very low, the average of the figures given above being 0.4 mm/day, in contrast with the evaporation along the coasts (see the result for Punta Preguntoiro) and in the interior where the mean potential evaporation is about 3.5 mm/day.

TABLE V. *Variation of the evaporation over the ria as calculated from Rohwer's formula.*

segment	$\bar{e}_s$	$\bar{e}_z$	$\overline{W}$	E
A	20.8	19.8	3.7	0.5
B	19.3	19.0	4.7	0.2
C	19.3	18.5	4.7	0.5
D	17.8	17.3	4.3	0.3
E	18.4	16.9	3.9	0.8
F	18.6	17.5	3.5	0.6
G	18.8	18.3	3.1	0.2
H	18.2	18.1	2.7	0.05

4. The oceanic area bordering the Galician west coast

4.1 Introduction

The oceanographic situation in the Rías Bajas of Galicia is influenced considerably by the conditions of the oceanic water bordering the west coast of Galicia. A study of the oceanography of these waters has to be made if one wants to investigate the conditions in the rias. A general survey is given of the oceanographic situation of this sea area, with the emphasis on the upper layers, down to about 200 m.

Various more or less detailed investigations have been published. A recent study of the water masses along the Spanish and Portuguese coasts is the one by MADELAIN (1967A). The following pattern is generally accepted.

Excluding the surface waters with more or less variable temperature and salinity, we may distinguish at first the North Atlantic Central Water, with a nearly linear temperature salinity relation. This North Atlantic Central Water extends from below the surface layers (that is from about 100 m) down to 500 or 600 m in depth. This water mass will be discussed in greater detail in the next section.

Below this depth there is a mixing zone with the Mediterranean Water, which is water flowing over the sill of the Straits of Gibraltar into the Atlantic. This water from the Mediterranean is characterized by high salinities and relatively high temperatures. The core of the Mediterranean Water (or, perhaps better: Gulf of Gibraltar Water, as at some distance from the Straits it is already mixed with the overlying water and therefore no longer has the properties of pure Mediterranean Water) is found at approximately 1000 m deep and has σ_t values between 27.7 and 27.8. A major vein of this water is found along the continental slope of the Iberian peninsula. Relatively large variations are found in the extension of this water mass both in the horizontal and in the vertical direction. Close inspection has revealed the existence of sometimes different cores at different levels (ELLETT, 1963).

At greater depths, that is below about 1.500 m the presence of North Atlantic Deep Water, mixed with smaller or larger quantities of Mediterranean Water, can be observed. As we are only concerned with the upper water masses, which exert an influence on the situation in the rias, a further discussion of this water is less relevant in connection with our subject.

Surface currents estimated by merchant vessels using the deadreckoning method are presented in a number of Maritime Atlases (e.g. NAVAL OCEANOGRAPHIC OFFICE 1965). They show rather persistent southerly currents with speeds of about 0.5 knots (25 cm/s). Information on sub-surface currents is mainly indirect, i.e. based on conclusions drawn from the distribution of the water properties and the results of dynamic computations of geostrophic currents. Only a small number of current measurements is known from sub-surface depths.

SAELEN (1961) reports on current measurements made in July 1958 on the Galicia

Bank at 42° 41′ N, 11° 47′ W. At a depth of 10 m an average current of 10 cm/s in a western direction was observed; at a depth of 100 m the current was 6 cm/s towards the southwest and at 600 m 3 cm/s towards the north-northwest. Current measurements made by LACOMBE (1961) in the same month at a position near 42° 34′ N, 10° 24′ W show a current of about 5 cm/s to the north at a depth of 1,000 m. The current at 10 m shows strong tidal influences. The residual current at that depth appears to have been in a northwestern direction, but the velocity is not given, probably because of the mediocre quality of the measurement and the difficulty to eliminate the tidal current.

MADELAIN (1967[B]), who has studied the geostrophic currents from observations of 1957 and 1958 and who has compared the results with G.E.K. observations (current data from measurements with a 'geomagnetic electro-kinetograph') has found indications of a northern current near the surface. However, he states that the current data obtained are complex and there can be some uncertainty as regards the validity of geostrophic calculations in this situation.

Another recent study, by FERNÁNDEZ (1970), reveals a rather complicated pattern of mostly weak relative geostrophic currents (with respect to the 600 db level) for the months October-November.

The oceanic area discussed here has several points in common with other areas along the east coasts of the oceans. It is clarifying to compare the characteristics given above and those that will be given in the following sections with the general description of eastern boundary currents by WOOSTER AND REID (1963).

4.2 The North Atlantic Central Water

The North Atlantic Central Water (NAC water) is a water mass that extends over large areas of the North Atlantic Ocean (SVERDRUP ET AL. 1942). ISELIN (1939) describes its formation by sinking of surface water in late winter in the middle of the North Atlantic. So there is an annual renewal of this water mass.

The average temperature-salinity relation for this water mass according to Sverdrup is a straight line in the t-S diagram through the points $S = 35.10‰$ $t = 8$ °C and $S = 36.70‰$ $t = 19$ °C. COOPER (1952) finds for the water of the Celtic Sea (south of Ireland) the relation $S = 35.508 + 0.124\,(t\text{-}11)$, with a range of temperatures between 10° and 11.5 °C. Cooper's temperatures are about 0.2 °C higher for the same salinities than Sverdrup's data. A superficial inspection of the available serial data for the sea region off the Galician coast shows that the observed temperatures and salinities are arranged in the t-S diagram in a zone along the lines given by Cooper and Sverdrup, but the scatter of the points in the t-S diagram is such that it does not appear justified to calculate another relation for the NAC water in this region. Because of the way this water mass is formed, one should not be surprised to find not only different relations for different regions, but also for different years. Evidence has been given by

WORTHINGTON (1954) that long-term variations of the properties of this water mass do occur.

4.3 Surface water

The surface layers are directly influenced by the exchange of heat and water with the atmosphere. They are subject to variations because of variable currents and variations in the mixing by eddy diffusion. Because of these variations, which often are much larger than those that occur at greater depths, the study of the surface layers has to be concentrated on the most important aspects. The large number of surface observations of temperature and, be it to a lesser extent, also of salinities, compared with data of deeper layers, makes an investigation of the average temperature and salinity the most obvious.

Recent data on the surface temperature of this region have been presented by BULLIG AND BINTIG (1954). For the surface salinity we have the data of BÖNHECKE (1936). For the present study these data were supplemented with data from the 'Bulletins Hydrographiques' till 1956 (CONSEIL INTERNATIONAL POUR L' EXPLORATION DE LA MER). The annual variation of temperature and salinity in the one degree square 42°-43° N, 9°-10° W according to these data is given in fig. 4.1.

The salinity graph shows irregular variations. As the averages for some months are biased towards certain years with a high number of observations, these variations do not necessarily represent a real feature of the average annual variation. In order to show clearly the annual variation and to eliminate accidental variations the salinity has also been averaged for overlapping three-monthly periods.

The annual variation of the temperature of this square does not show conspicious features. However, if a smaller area, near the coast is considered, a quite different pattern is observed, as has been shown by ANADÓN ET AL. (1961). The annual variation of temperature near the entrance of the Ría de Vigo, as given in the publication of these authors, is also presented in fig. 4.1. The temperatures near the entrance of the Ría de Vigo do not differ much from those of the one-degree square from November till June. However, in summer, and especially in July, the temperatures are much lower. These low temperatures are known to be caused by upwelling. The reason why in the larger area before the coast there is not so much indication of upwelling apparently is that this phenomenon occurs locally, and that the major shipping routes cross the area outside the main regions of upwelling.

4.4 Near-surface layers

The near-surface layers form a transition between the surface water and the deeper waters, which have more or less constant values of temperature and salinity.

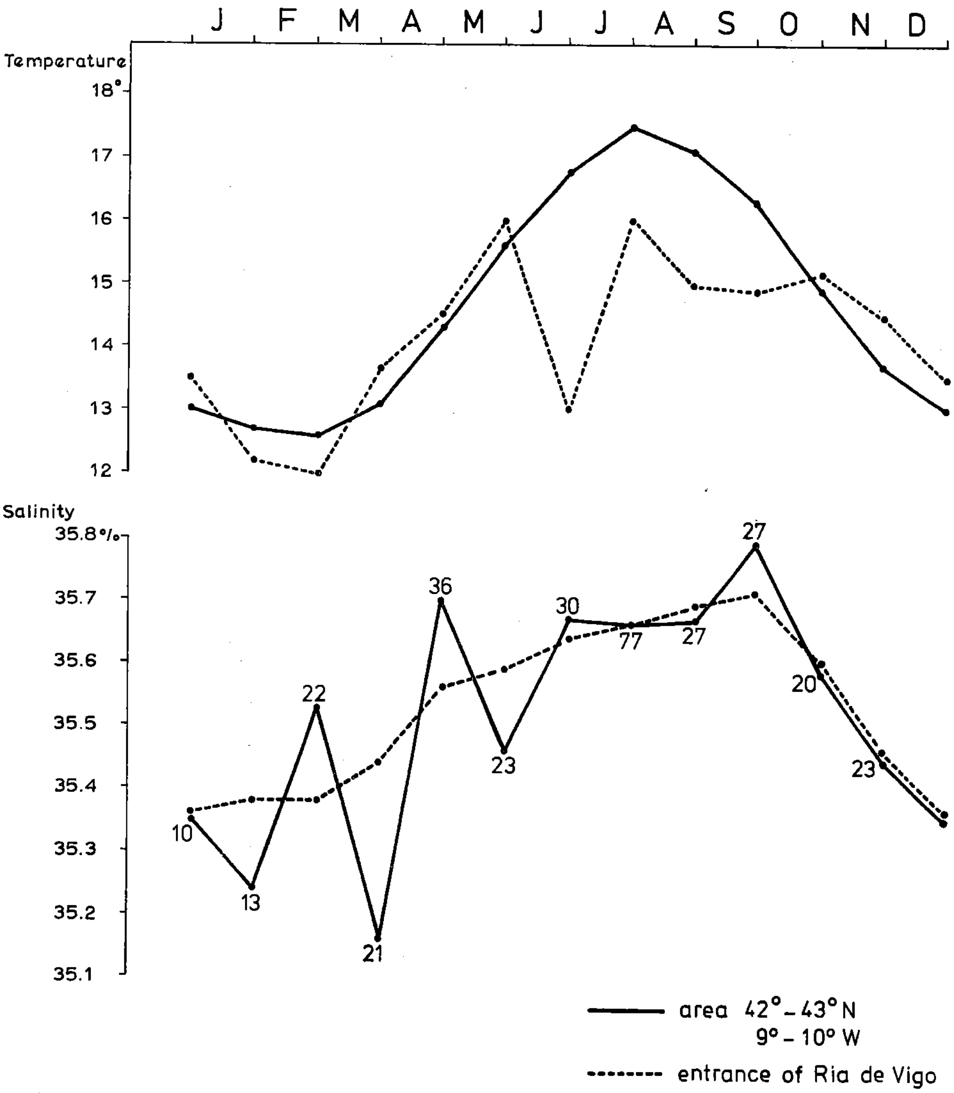


Fig. 4.1 Annual variation of surface temperature and salinity at sea in front of the Ría de Arosa and of the temperature at the entrance of the Ría de Vigo.

In order to study the annual variation of the properties of these layers data from different sources have been collected. The area considered is the area between 41° 45′ N and 43° 15′ N, and between the meridian of 10° W and the coastline, excluding the rias. In Table VI a list is given of the data used.

TABLE VI. *Serial Observations, area 41°45′ N-43°15′ N; east of 10° W*

January	1934 Théodore Tissier	1 stat.	Rapp. et Proc. Verb. *93*
	1950 Théodore Tissier	2	Bull. Hydrogr.
	1962 Labor. Invest. Pesq. Vigo	1	Inv. Pesq. *31* (1)
	1963 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
February	1928 Eduardo Dato	2	Rapp. et Proc. Verb. *62*
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
March	1930 Xauen	3	Rapp. et Proc. Verb. *72*
	1958 Michael Lomonosov	1	Trud. Morsk. Hidrof. Inst. *21*
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
April	1910 Michael Sars	1	Rep. Results M. Sars
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
May	1923 Tanche	2	Rapp. et Proc. Verb. *35*
	1925 Marques de la Victoria	3	,, *40*
	1926 Proserpina	3	,, *44*
	1927 Eduardo Dato	2	,, *55*
	1927 Albacora	1	,, *55*
	1930 Xauen	3	,, *77*
	1949 Théodore Tissier	2	Bull. Hydrogr.
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
June	1922 Dana	1	Hydr. Results Dana Exp. I
	1928 Eduardo Dato	2	Rapp. et Proc. Verb. *62*
	1929 Laya	4	,, *70*
	1930 Dana	1	Hydr. Results Dana Exp. I
	1932 Xauen	3	Rapp. et Proc. Verb. *87*
	1952 Théodore Tissier	2	Bull. Hydrogr.
	1955 Théodore Tissier	7	,,
	1956 Théodore Tissier	7	,,
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
July	1922 Tanche	3	Rapp. et Proc. Verb. *31*
	1924 Tanche	1	,, *37*
	1950 Théodore Tissier	1	Bull. Hydrogr.
	1958 Calypso	2	Cahiers Océanogr. *12*
	1962 Labor. Invest. Pesq. Vigo	3	Inv. Pesq. *31* (1)
	1965 Snellius	1	manuscript

The region was divided into two parts, one east of the 200 m isobath, and the other west of this line. These two regions are representative of the shelf and of the continental slope, respectively. The monthly or bi-monthly averages of temperature and salinity were calculated for a number of depths. The bi-monthly averages were calculated for the winter months, where the number of obesrvations is relatively small. The thus obtained annual variation at different depths is shown in fig. 4.2.

The temperatures at 50 m and 100 m in the slope area show a normal variation. If we compare them with data of HELLAND HANSEN (1930) for the Bay of Biscay we find only a larger range of the annual variation, as is shown in the following table:

August	1926 Proserpina	3	Rapp. et Proc. Verb. *44*
	1927 Eduardo Dato	3	,, *55*
	1932 Xauen	3	,, *87*
	1949 Théodore Tissier	2	Bull. Hydrogr.
	1951 Xauen	8	Bol. Inst. Esp. Oceanogr. *83*
	1952 Xauen	11	,, *103*
	1953 Xauen	4	,, *103*
	1954 Xauen	39	,, *103*
	1957 Calypso	2	Cahiers Océanogr. *12*
	1959 Amiral Mouchez	1	,, *13*
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
September	1930 Xauen	4	Rapp. et Proc. Verb. *77*
	1936 Albacora	4	Bull. Hydrogr.
	1953 Xauen	1	Bol. Inst. Esp. Oceanogr. *103*
	1954 Xauen	3	,, *103*
	1958 Discovery	2	Woods Hole Oc. Inst. ref. 59-54
	1962 Labor. Invest. Presq. Vigo	2	Inv. Pesq. *31* (1)
October	1955 Aventure	1	Bull Hydrogr.
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
November	1925 Marques de la Victoria	3	Rapp. et Proc. Verb. *40*
	1933 Théodore Tissier	2	,, *93*
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)
December	1923 Cinco de Outubro	2	Rapp. et Proc. Verb. *35*
	1927 Eduardo Dato	2	,, *55*
	1932 Challenger	1	Bull. Hydrogr.
	1962 Labor. Invest. Pesq. Vigo	2	Inv. Pesq. *31* (1)

Rapp. et Proc. Verb. = Rapports et Procès-Verbaux des Réunions, Conseil Perm. Internat. Explor. de la Mer
Inv. Pesq. = Investigacion Pesquera

	annual temp. range		months of temp. maximum and minimum	
	Bay of Biscay	Slope area	Bay of Biscay	Slope area
50 m	2.4°	4.0°	Sept.-Oct./March	Sept.-Oct./March-April
100 m	0.7°	1.2°	December/March	Nov.-Dec./March-April

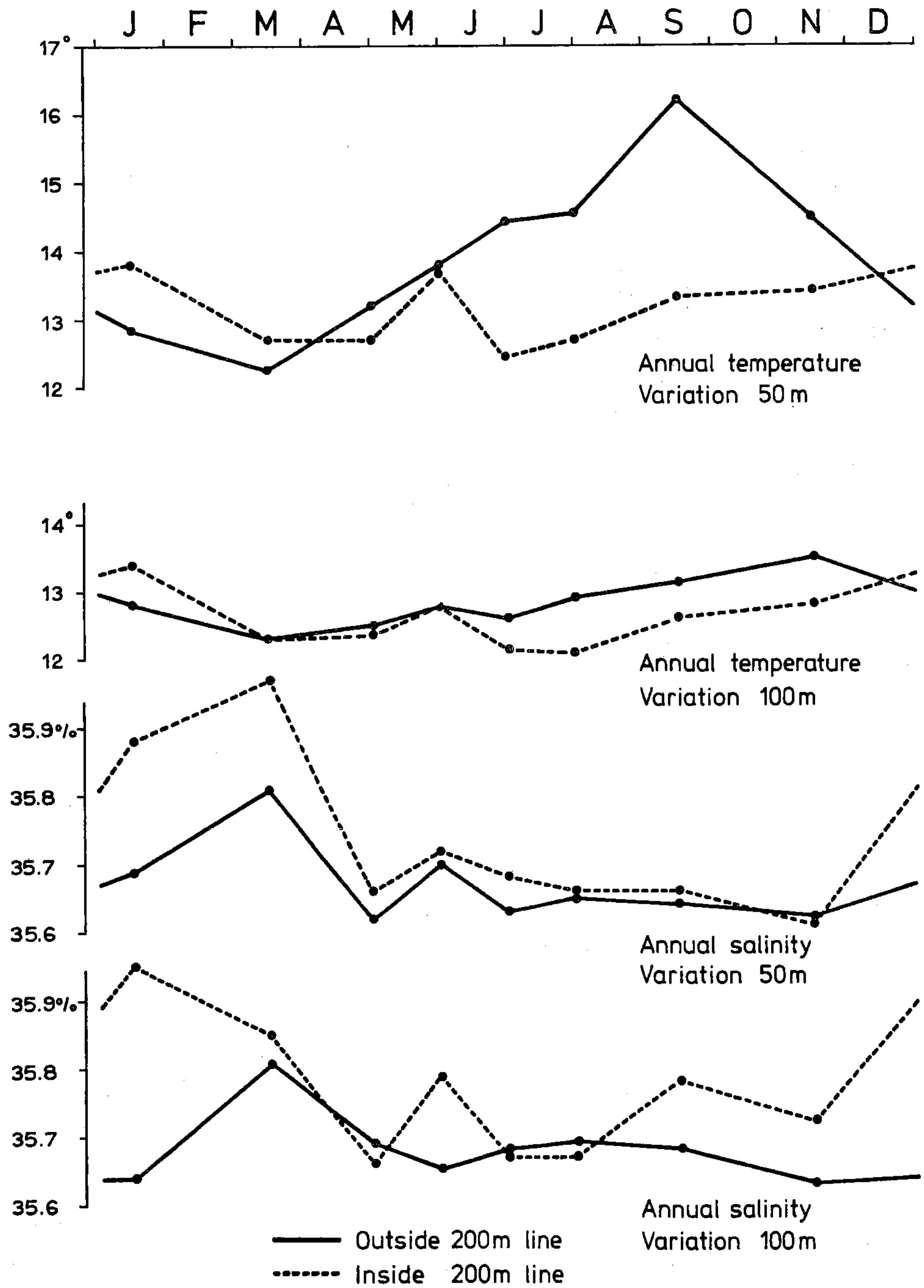


Fig. 4.2 Annual variation of salinity and temperature at different depths in front of the Ría de Arosa.

On the shelf the temperature shows a more complex variation. The temperatures in the beginning of the year are here slightly higher than further offshore, but the trend is the same as for the variation of the temperature in the slope area. This changes between June and July. Then the temperatures drop and reach minimum values. This apparently is caused by the upwelling of colder, deeper water, the effect of which is most pronounced in July and August. During the autumn and early winter the temperature increases gradually, till the maximum is reached in January-February.

The most conspicuous feature of the annual variation of the salinity is the occurrence of high values in winter, especially over the shelf. These high salinities appear to be limited to the sub-surface layers, as at the surface we do not find such high values.

In fig. 4.3 the average temperature profiles of a section perpendicular to the Galician coast are given for different months. These figures show that in March-April the water appears to be practically isothermal, and that a gradual warming-up of the surface layers occurs in May and June. The profiles for the following months clearly show the upwelling in summer. Even the November-December profile shows an uplifting of the isotherms towards the coast, which suggests that this proces of upwelling may still be active.

Whether this sequence of events is nearly always the same for every year is not certain. A series of observations made in 1971 on a section WNW of la Coruña strongly suggests the initiating of the upwelling process already in the end of April (OTTO, 1972).

The upwelling water for a considerable part consists of North Atlantic Central Water, as may be seen from fig. 4.4, in which the average values of the temperature and the salinity of the shelf stations have been plotted in the t-S diagram. The 100 m values in part fall within 0.05 $^0/_{00}$ from the t-S relation for N.A.C. water according to Sverdrup, some values being only just outside this zone. Only the March-April values differ considerably, as is also the case with the 50 m values. The other 50 m values are to the left of Sverdrup's line in the t-S diagram; this probably reflects the modification of N.A.C. water by warming-up or by admixture of low-salinity water. The upwelling is an irregular phenomenon, as is also the case in other regions of upwelling. This irregularity may be illustrated by the results of the observations from the Spanish research vessel 'Xauen', obtained from 13-16 August 1954. In fig. 4.5 the depth of the $\sigma_t = 27.0$-surface is shown. The irregular shape of the isobaths of this surface may be either a real picture of the synoptic situation at that time, or it may be the consequence of short-time variations of the depth of this surface within the time intervals between the successive stations. Probably both effects act simultaneously.

The variability of the upwelling is also demonstrated by the frequency distribution of the bottom temperatures on the shelf. In fig. 4.6 this frequency distribution is given for the month August, as calculated from the observations mentioned in table VI.

Further data on the upwelling in this region can be found in a study by MOLINA (1972).

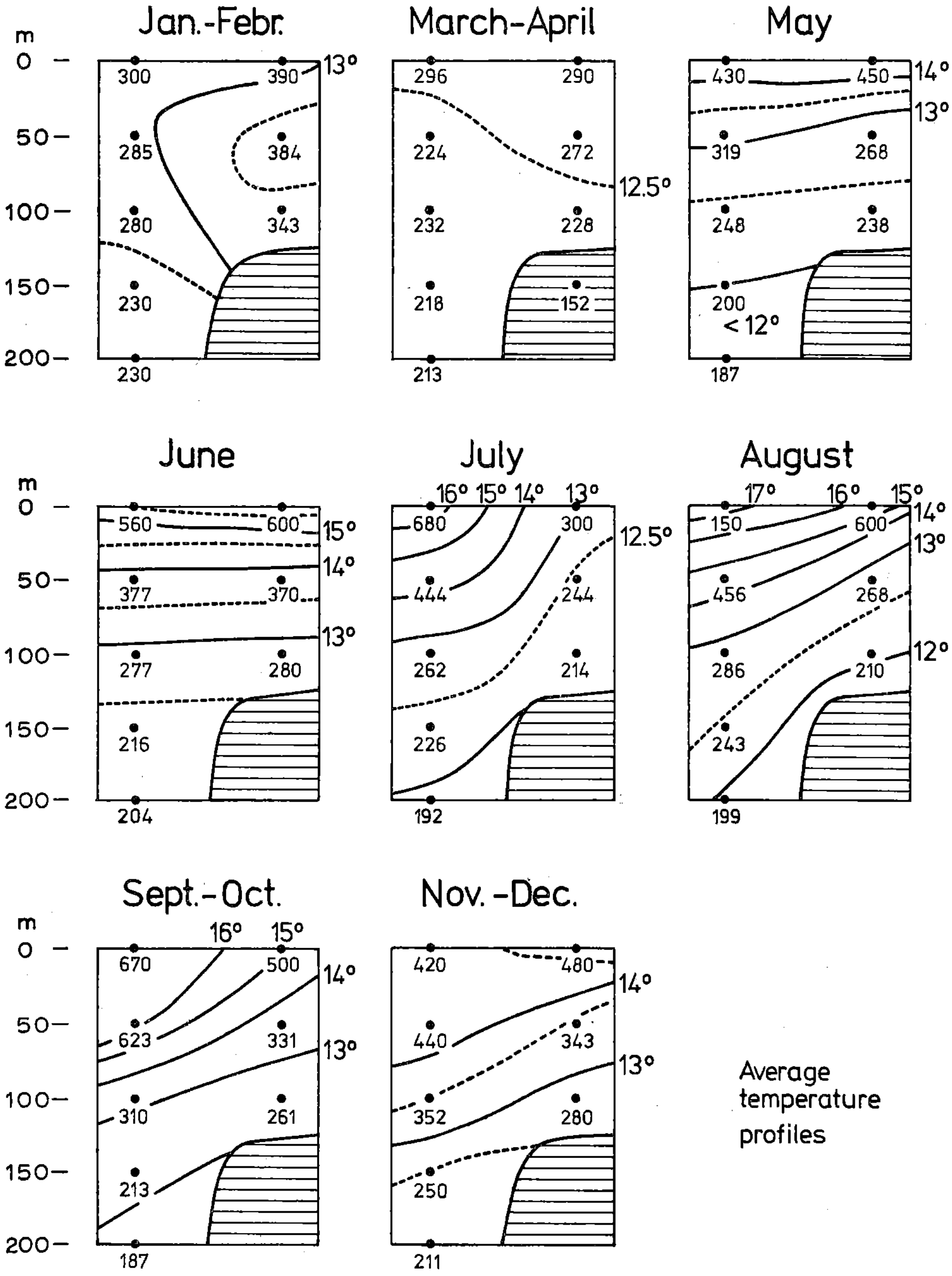


Fig. 4.3 Average temperature profiles in a section perpendicular to the Galician coast.

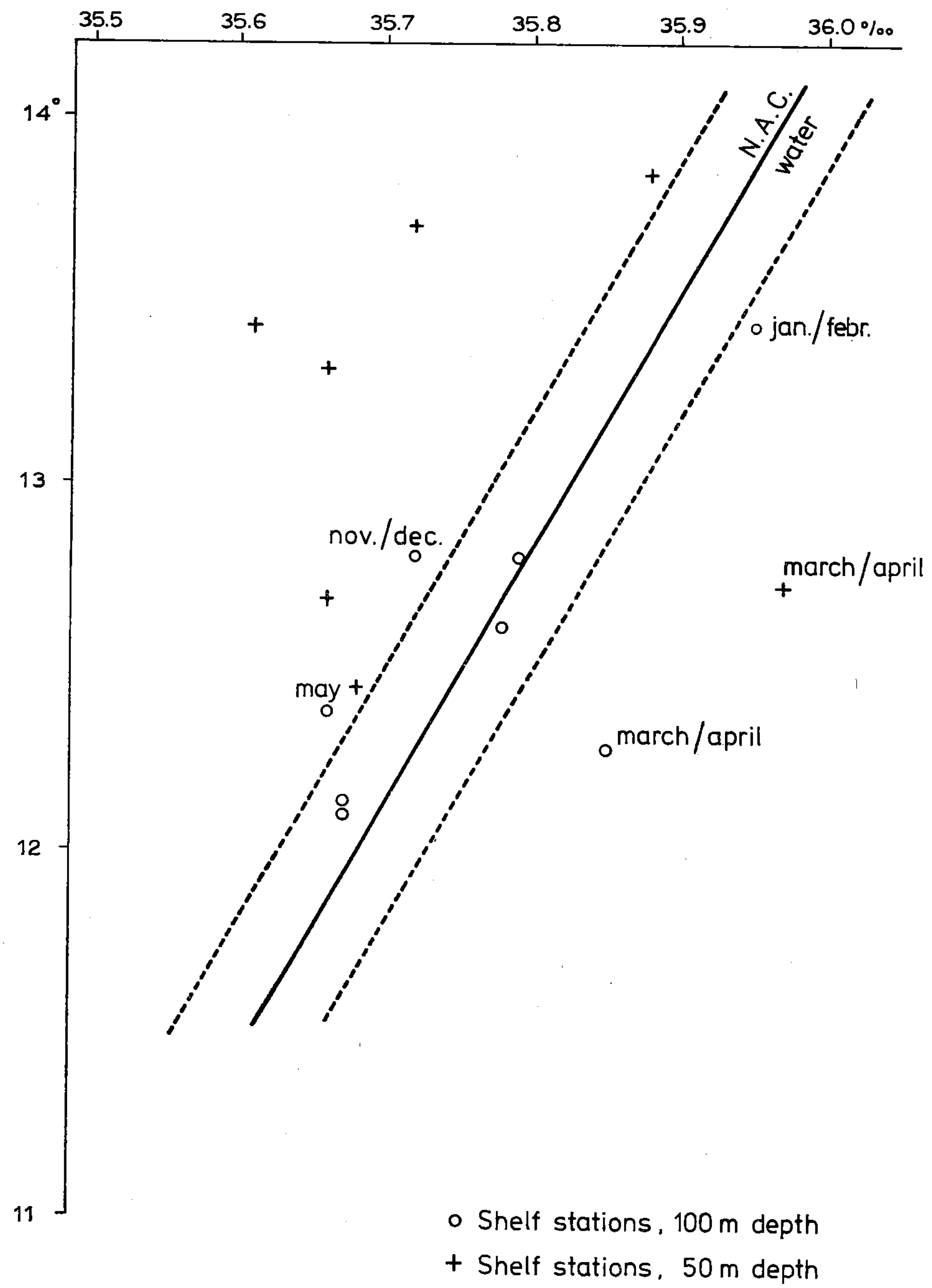


Fig. 4.4 *t-S* diagram for shelf stations off Galicia (monthly or bi-monthly means).

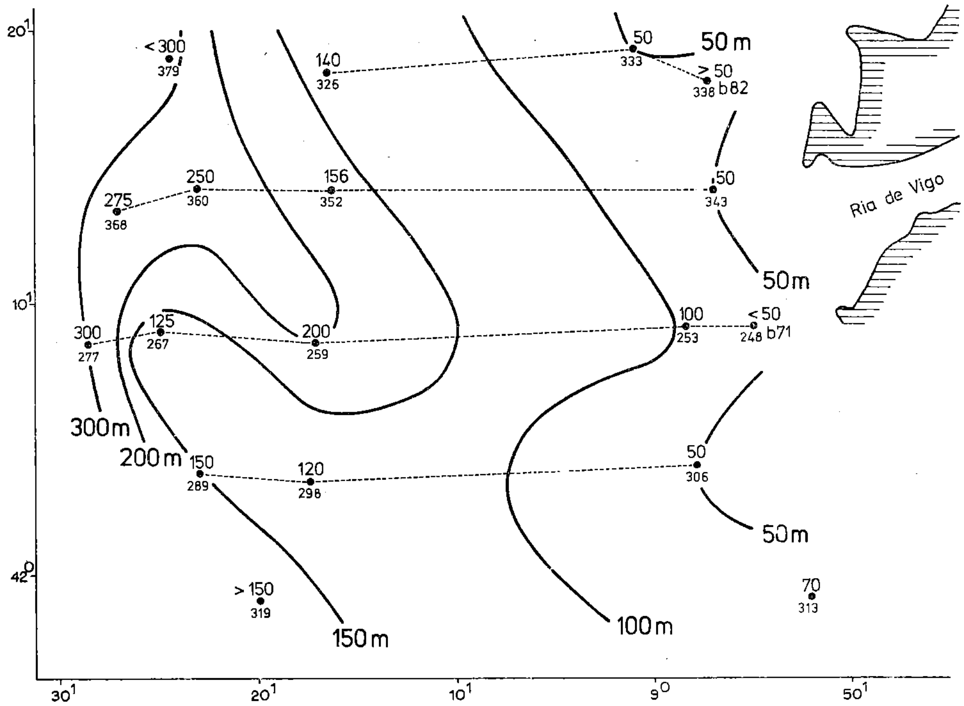


Fig. 4.5 Depth of σ_t = 27.0 surface, 13-16 August 1954.

4.5 The high-salinity water

During spring occasionally water of high salinity is observed at small depths. By this we mean water with salinities that are higher than the salinities of N.A.C. water of the same temperature. Often salinities of more than 36 $^0/_{00}$ are observed in this water mass. This water was observed on several ooccasins in the Ría de Arosa (see section 6.3). A conspicuous case was reported by FRAGA (1967) in the year 1962; it occurred in front of the Ría de Vigo. This water, according to Fraga, appeared to form a thin layer which ascended from a depth of 100 m in March till 20 m in June.

In March a salinity as high as 36.43 $^0/_{00}$ was observed with a temperature of 13.0 °C.

The occurrence of high-salinity water is also manifest from the March-April points in the *t-S* diagram of fig. 4.4.

A special case, when a water layer of very high salinities was observed, was in June 1955, during a cruise of the French research vessel 'Théodore Tissier''. The high-salinity water at that occasion did not extend to over the continental shelf, but further offshore values up to 36.5 $^0/_{00}$ were found. The temperatures of the water with a salinity of more than 36 $^0/_{00}$ were between 13 °C and 14 °C (see fig. 4.7).

About the origin of this water we are not quite sure. An upwelling of Mediterranean

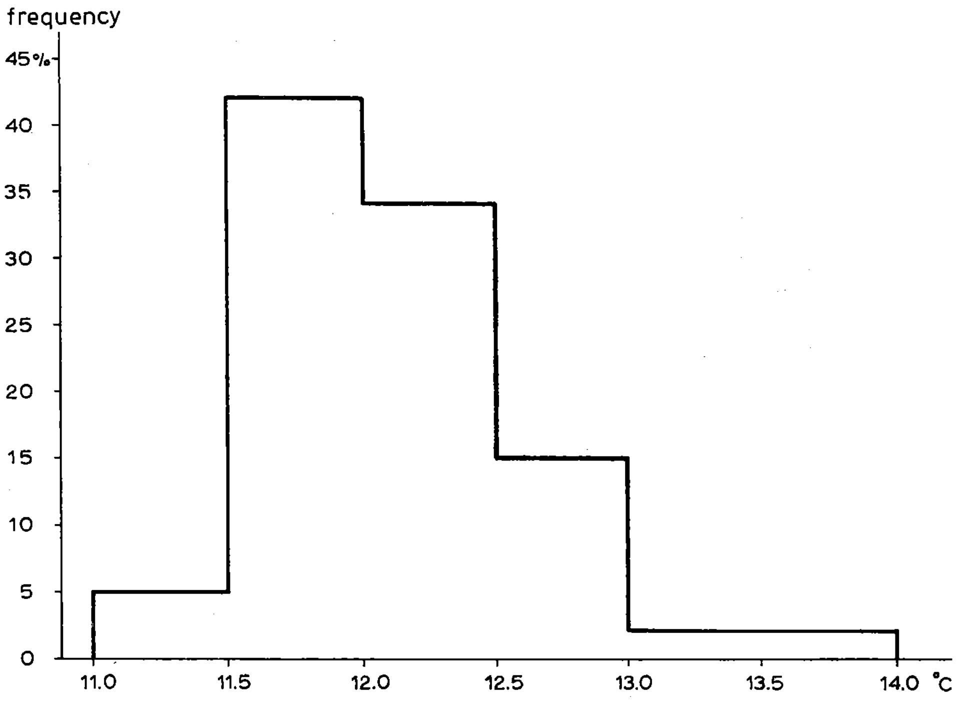


Fig. 4.6 Frequency distribution of bottom temperatures on the shelf (August).

water is out of the question, as, apart from the improbability of upwelling from depths of 800 m and more, the observed temperatures and salinities do not fit with the *t-S* values of the Mediterranean water mass.

The data available are too few to make possible a definite conclusion about the source of the high-salinity water. However, it is thought likely that it originates from the sea surface further offshore. At about 11° West water of the right salinity and temperature is found from December till April (BÖHNECKE, 1936). The water is probably also related to the 'high-salinity pockets' that have been described by MC.BARY (1963) for the Bay of Biscay. It is not clear whether the occurrence of the high-salinity water also coincides with the phenomenon in the older oceanographic literature known as the 'transgressions Atlantiques' (see for instance LE DANOIS, 1925).

4.6 Concentration of nutrients in the ocean water

The concentrations of nutrients and of dissolved oxygen show an annual variation owing to variations of the biological activity and of the changing circulation pattern.

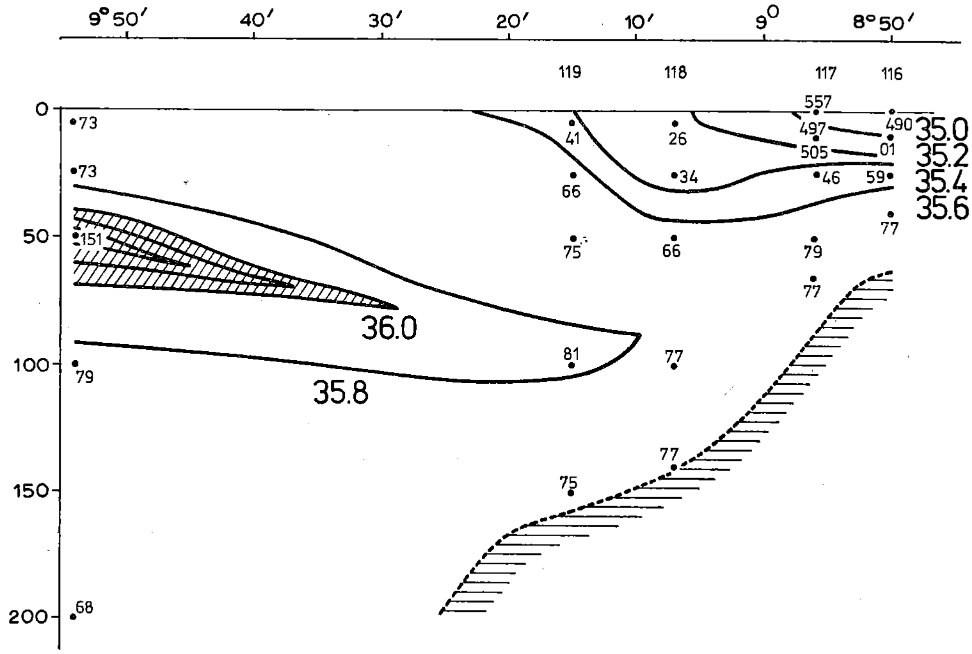


Fig. 4.7 Presence of high-salinity water off Galicia, June 1955.

These variations are illustrated by the results of some serial observations made at different times of the year. These results are presented as graphs of concentration versus depth in fig. 4.8, together with the data on salinity and temperature. The data are from observations by the 'Michael Lomonosov' and the 'Discovery II' in 1958 at about the same position in March and September respectively, and by the 'Dana' in June 1930, somewhat more to the south. The data on phosphate-phosphorus (PO_4-P) of the Dana have been used as corrected by COOPER (1938).

In March the surface layers down to a depth of 100 m are practically homogeneous, while towards greater depths the density increases only gradually. Apparently a good vertical mixing of the surface layers is accomplished by the high turbulence, caused by the wind, and by convection, caused by cooling at the surface. There is a slight undersaturation of oxygen in the first 100 m of water and the phosphate-phosphorus content is relatively high, indicating a low level of phytoplankton production.

In June, at the beginning of the upwelling season, the water is stratified and the oxygen saturation at the surface is 100%. The anorganic phosphorus is exhausted down to 75 m of depth. This suggests high level of phytoplankton production in the preceding period.

In September the stratification of the water near the surface is still more pronounced than in June. The content of phosphate-phosphorus however, is higher, which may be caused by regeneration from organic phosphorus or by advection.

The annual variation of the various nitrogen compounds has been investigated by FRAGA (1967) at position on the shelf in front of the Ría de Vigo. His figures show high surface values of the concentration of nitrate in winter (NO_3-N values up to 11.3 μgat/litre in January) and generally low surface values (down to 0.4 μgat/litre) in the summer, in accordance with the annual cycle of plankton production. However, the deeper layers show maximum values of nitrate in summer (up to 13.6 μgat/litre at 100 m in August, and occasionally high values occur even at the surface, an effect that can be considered to be caused by upwelling.

Other nutrients will probably show a similar pattern of the annual variation, as for instance is indicated by measurements in 1958 of silicate by the 'M. Lomonosov' and the 'Discovery II'.

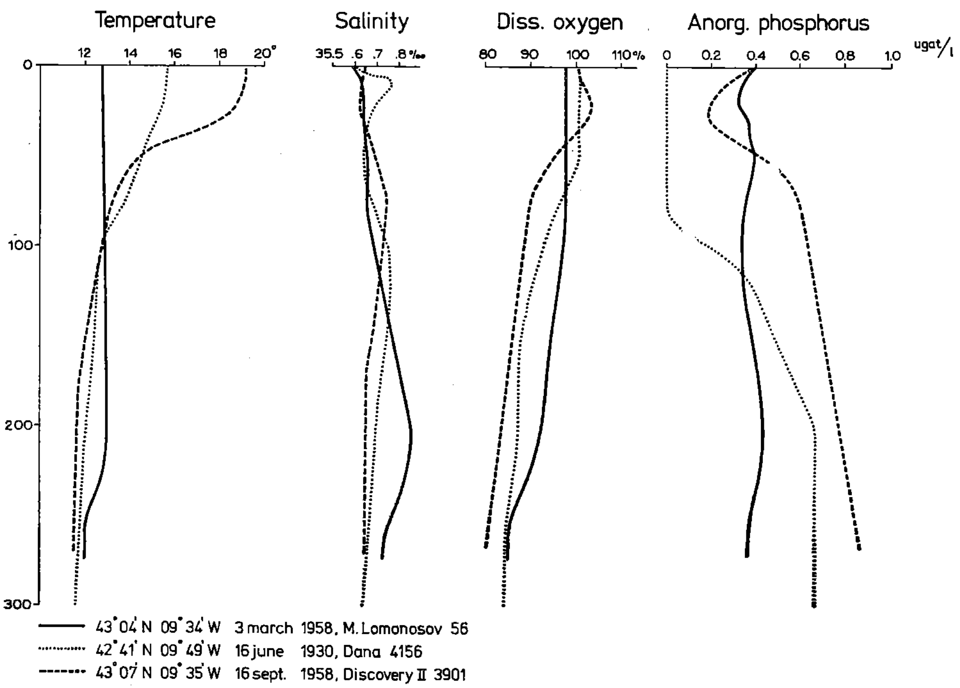


Fig. 4.8 Examples of vertical distribution of nutrient concentrations off Galicia.

5. Methods of observation and analysis

5.1 Observational programme at Punta Preguntoiro

At the base camp at Punta Preguntoiro the following programme of observations was carried out:

1. Meteorological observations (see section 2.3).
2. Observations of water temperature and sampling for salinity determination.
3. Visual estimation of wave height (summer 1964 only).
4. Sea level observations with a recording tide gauge.

The meteorological observations need no further comment here. Water temperature was observed and salinity samples were taken at the end of the landing pier, using a plastic bucket. For temperature measurements a sea water thermometer was used with an accuracy of 0.1 °C.

The accuracy of the visual wave height estimates cannot be considered to be high in this case where different generally unexperienced observers had to do this job. They just serve to give an indication.

The tide gauge was provided by the Netherlands 'Rijkswaterstaat' (Department of Transport and Public Works). It consisted of a vertical galvanized tube with a free acces of the water at the bottom. On top of this tube a clock-driven recorder was placed on which, by means of a system of a wire and a pulley, the position was recorded of a float that moved up and down in the tube with the water level.

The major problem with this instrument was to ensure the accurate vertical position of the tube, as otherwise the float could not move freely. It proved very difficult to provide for a construction for fastening of the tube with sufficient rigidity to withstand wave attack during bad weather and shocks from colliding ships. Therefore no continuous series of good quality recordings could be obtained, and as the position with respect to standard level was not determined and furthermore changed somewhat with each re-installation of the gauge, the recordings cannot be referred to an absolute level.

5.2 Shipboard observations

Different ships were used for work on the ria, as mentioned in section 1.4. For different types of observations and for sampling a small hand-driven winch with a boom was used that could be fastened on the board of the ship.

The positions were determined from two simultaneously measured angles between three landmarks (lighthouses, beacons, mountain peaks). Under favourable circumstances this could result in an accuracy of the position of some tens of metres. Haze, rough weather and the absence of convenient landmarks sometimes proved to be serious impediments for a good position finding. In those cases additional information from depth observations could be helpful.

In 1962 and 1963 depth was determined by means of a small portable echo-sounder (made Pye Medway), and also by paying out the oceanographic wire till the weight touched the bottom, which could be observed from the tension in the wire. In 1964 a number of echograms was obtained with a recording echo-sounder that was lent by the Royal Shell Exploration and Production Laboratory (see KOLDIJK, 1968).

5.3 Observations of temperature

Surface temperature was measured with a sea water thermometer with an accuracy of 0.1 °C. Serial measurements were made with one or two reversing thermometers (Richter und Wiese) on a Nansen bottle. Temperature corrections were applied according to the normal oceanographic procedure (see U.S. HYDROGRAPHIC OFFICE, 1955). The estimated accuracy is 0.02 °C.

Temperature registrations were obtained with a bathythermograph. The temperature readings from the thus obtained BT slides have an accuracy of about 0.1 °C. The accuracy of the recorded depth is about 0.3 m.

5.4 Salinity determination

Salinity samples were taken by bucket (surface samples) and with Nansen bottles. A modification of the Knudsen method for chlorinity titration was used as described by VAN DAM (1940), using Copenhagen standard sea water as reference. Salinity was calculated using the well-known relation

$$S = 0.03 + 1.8050\, Cl$$

Within the limits of accuracy this relation gives the same results as the recently accepted redefinition (WOOSTER ET AL. 1969). The accuracy of the salinity values can be estimated to be about 0.05 $^0/_{00}$.

5.5 Secchi-disc observations

The Secchi-disc is a simple instrument for the determination of the light attenuation in the water. The results are influenced by the circumstances of observation, and therefore are not easily related to a well-defined physical property. A detailed account of the meaning of Secchi-disc data has been given by TYLER (1968).

The Secchi-disc used was a 30 cm, white painted disc. In the shallow water of the rivers the 'Secchi depth' often exceeded the bottom depth. Therefore in 1964 the brownish painted underside of the disc was used, giving smaller depths. For further details on these data the reader is referred to section 8.6.

The accuracy of the Secchi depths within the range of values observed here is about 0.5 m.

5.6 Observations of suspended matter

The sampling procedure used for the determination of the concentration of suspended matter was the one described by POSTMA (1954). The concentration was determined by filtering the samples and then weighing the filters. With the filters used the amount of suspended particles down to a size of some microns could be determined. The accuracy of weighing and the variation in weight of the filters, due to changes in the moisture content of the filters, limit the accuracy of this method. It was estimated that an accuracy of about 0.5 mg/l was obtained in the concentration values. As the filters were dried before weighing the figures refer to the weight of dry particles.

5.7 Observations of dissolved oxygen and nutrients

The determination of dissolved oxygen was done according to the well-known Winkler method (WINKLER 1888, JACOBSEN ET AL. 1950). Fixation of the samples was done directly after sampling aboard the ship. Storage on the ship, however, proved difficult. Precautions were taken against too large variations in temperature of the samples, and the results were critically examined for possible erroneous data. The accuracy of the accepted data was estimated to be no worse than 0.05 ml/l. Saturation values were calculated according to CARPENTER (1966).

Inorganic phosphorus, nitrate and nitrite were determined using the methods of HARVEY (1955) for phosphate, MULLIN AND RILEY (1955) for nitrate and WATTENBERG (1937) for nitrite. The estimated accuracy of the analytical procedures is:

phosphate	0.03 μg-at/l
nitrate	0.3 μg-at/l
nitrite	0.03 μg-at/l

As will be mentioned in paragraph 10.1 there is, however, uncertainty whether in some occasions the total error is not significantly greater than the analytical error, because of difficulties in storage of the samples during the cruises.

5.8 Observations of dissolved organic carbon

A small number of samples was taken for measuring the concentration of dissolved organic carbon. Analysis was performed according to the method given by DUURSMA (1960). The analysis had to be done at the Netherlands Institute for Sea Research, which meant storage and transport to the Netherlands. Samples were taken at the end of the campaigns in order to keep the period of storage as brief as possible. For the North Sea Duursma has found a reduction of 10 % of the dissolved organic carbon values after storage over 1 or 2 months.

Contamination of the samples by grease of the Nansen bottles cannot be ruled out completely, but is thought to be small because of the intense use of the bottles after greasing and Duursma's experiences under similar conditions.

Apart from the effects of storage and sampling the accuracy is determined by the analytical method which is about 0.03 mg C/kg.

5.9 Current measurements

Various methods for current measuring were used. In 1962 a pendulum-type current meter was employed; the inclination of a pendulum, which is suspended in the water, indicates the current velocity.

In 1963 and 1964 an Ott current meter was used, in which the number of revolutions of an impeller is measured electrically aboard the ship. In both cases current direction was measured by noting the horizontal direction in which the measuring instrument was deflected by the current.

In 1963 and 1964 drifters of two different types were used as well. One consisted of ballasted bottles, just protruding from the water, marked with a small yellow flag, and from which at 1 metre depth a wooden current cross (30 × 30 cm) was suspended. The other consisted of yellow balls from which a current cross of 1 × 1 metre at a certain depth between 10 and 30 m) was suspended. The path followed by these drifters (usually for several drifters simultaneously) was determined by regularly tracing them and by determining their position as accurately as possible in the usual way.

Comparison of the drift over a fixed distance of drift bottles and current measurements with the Ott meter revealed a rather satisfactory agreement. The Ott meter gave velocities between 21 and 26 cm/s while the drifters gave velocities between 18.5 and 20 cm/s, a difference which may be the result of vertical variation of the current.

The accuracy of the measurements is influenced by the time interval over which the (mean) current was measured. However, an accuracy of about 2 or 3 cm/s, is thought to be a good estimate for most of the observations.

5.10 Quality control

The large numbers of observations and samples, the difficulties encountered in storage of samples, the improvisations that sometimes were necessary in the conditions under which the work had to be done and the fact that the work had to be carried out by partly inexperienced, be it devoted, observers, may all have caused smaller or larger errors in the different determinations. We are, however, confident that no major errors are present in the data that finally were accepted after scrutinizing them carefully and rejecting data that for one reason or an other appeared to be doubtful.

For temperature and salinity data a useful check was a control of the downward increase of the density calculated from these data. The vertical trend of oxygen, phosphate, and nitrate normally is a decrease of the first property and an increase of the two latter properties with depth. This rule, however, has its exceptions and before rejecting data other factors had to be taken into consideration.

Of course such a quality control is more or less subjective. By no control at all oddities are introduced in the basic observational material that have no relation with reality. Too rigid control procedures produce a picture of only those processes that do not contradict the preconceptions on which the control procedures are based. Here it was attempted to steer a middle course.

6. Oceanography of the Ría de Arosa in summer

6.1 Average distribution of temperature and salinity

During the three summer campaigns of 1962, 1963 and 1964 numerous observations of temperature and salinity were made. Taking together all these observations we get a picture of the average conditions during these three periods. As the weather conditions and the run-off were not so very different during these campaigns and did not deviate much from the averages for this time of the year (July and August), we may suppose that the average oceanographic conditions will reflect the normal summer conditions.

In order to get an impression of the regional distribution of temperature and salinity, some charts have been drawn, showing the average values of the observations. These averages were computed over small rectangles with dimensions of $2\frac{1}{2}$ nautical miles in the east-west direction and $1\frac{1}{2}$ nautical miles in the north-south direction, or, where steep gradients of temperature and salinity made this necessary, over smaller rectangles, of a half or a quarter this size. In nearly all these rectangles a fairly large number of observations from different years was available; only some of the rectangles in the northwestern bays contain a small number of observations or observations from only one year. However, a coherent pattern of the distribution of the water properties was obtained, which gives confidence in the charts being representative.

These charts show isolines of salinity and temperature at the surface and of salinity in the layer between 20-40 m depth. They are presented in fig. 6.1, 6.2 and 6.3.

The charts give an impression of some of the more important features of the circulation of the water in the ria.

Considering the distribution of the surface salinity as shown in fig. 6.1, we see a tongue of water with relatively low salinities protruding from the mouth of the Rio Ulla in a southwesterly direction, along the right (northwesterly) shore of the ria. This tendency of the outflowing, less saline water to follow the right side of an estuary in the northern hemisphere, is a well known effect of the Coriolis force. In the inner part of the Ría de Vigo (in the Ensenade de San Simón) the same has been observed by SAIZ ET AL. (1961).

The influence of the Rio Umia in the southeastern part is much smaller, and a well defined tongue of less saline water is not apparent.

The tongue of relatively high salinity water, protruding into the ria from the ocean in the direction of Isla de Arosa may be caused by an inflow at the surface or by upwelling or vertical mixing.

A remarkable point is also the presence of more saline water in the bays of Puebla de Caraminal and Rianjo in the northwest, where the salinity at the entrance of the bays is lower than at their head. The small quantities of fresh water advected by rivu-

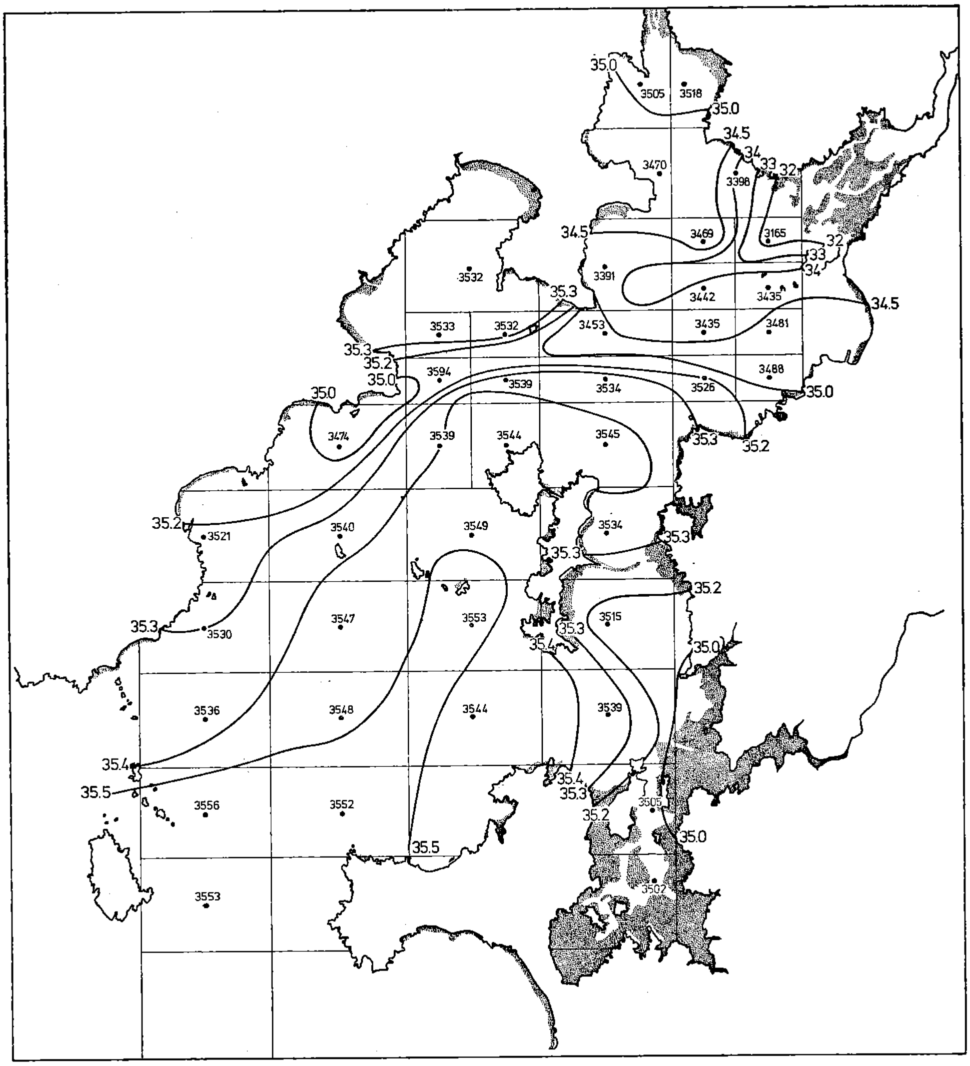


Fig. 6.1 Salinity distribution at the surface, Ría de Arosa, summer.

lets, the most important of which is the Rio Beluso, have no significant influence on the properties of the water in these bays.

The surface salinities observed in July 1968 by GÓMEZ GALLEGO (1971) are very similar to those presented in fig. 6.1, with one exception, in the bay of Rianjo much lower salinities occurred, indicating a larger influence of the rivers Ulla or Beluso in that part of the ria.

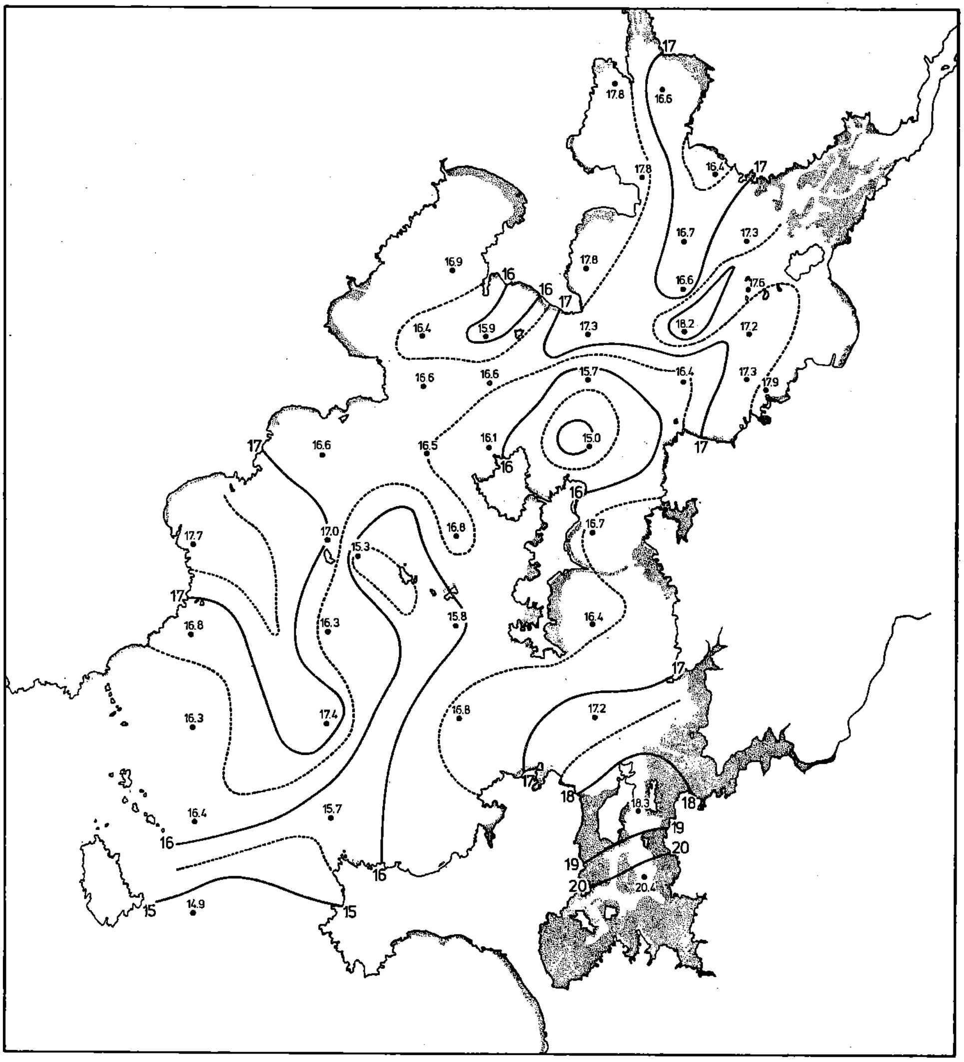


Fig. 6.2 Temperature distribution at the surface, Ría de Arosa, summer.

The temperatures at the surface (fig. 6.2) are the result of the warming up by heat exchange with the atmosphere and by insolation as well as the advection of and the mixing with warmer water from the rivers and colder water from the ocean or from sub-surface layers.

In various respects the pattern is therefore different from the pattern of surface salinities. But nevertheless the tongue of warmer water from the river along the right

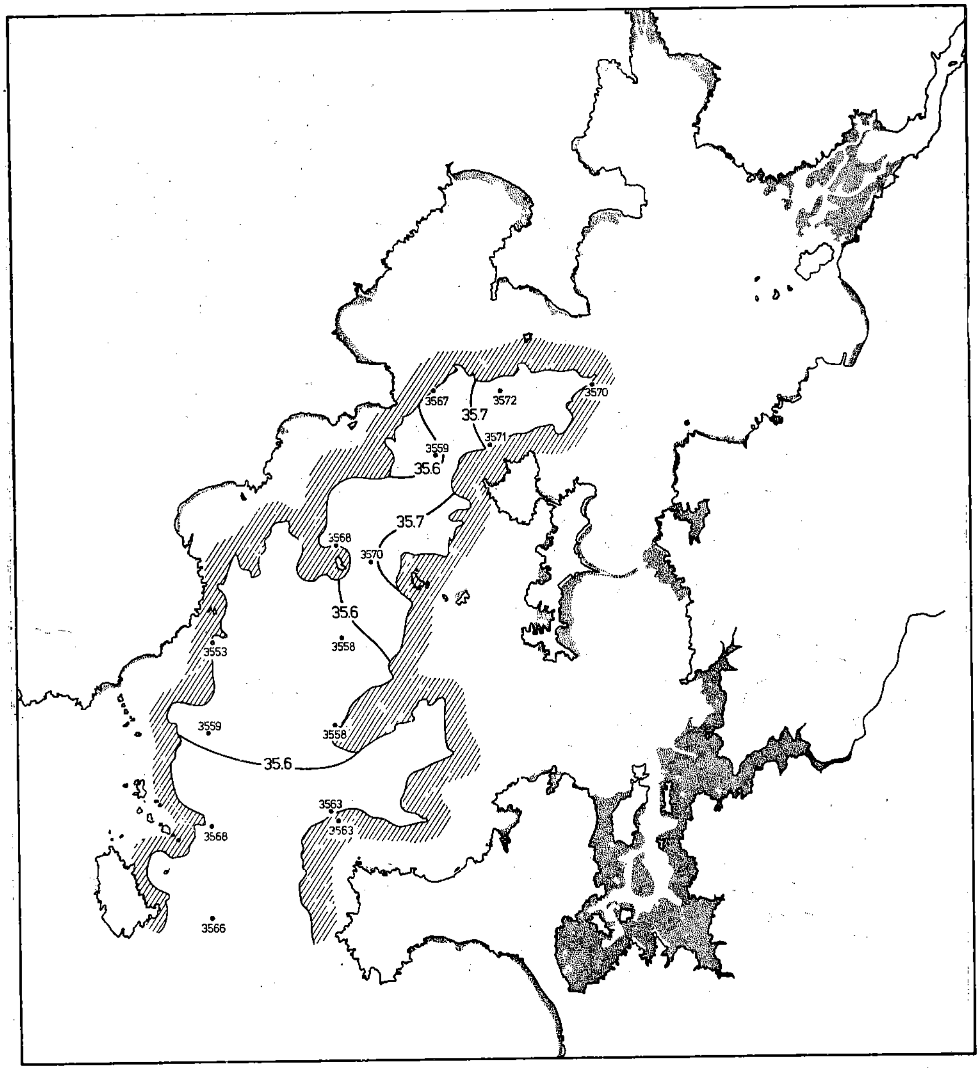


Fig. 6.3 Salinity distribution at 20-40 m depth, Ría de Arosa, summer.

shore is as prominent a feature of the temperature distribution as its low-salinity counterpart in the salinity distribution. The surface temperatures observed by Gómez Gallego in July 1968 are about 1.5 °C higher than the means for the summers of 1962-1964. However, the relative distribution appears to be similar. In fig. 6.3, only small variations in the average salinity in the layer between 20 m and 40 m depth can be seen. It is remarkable that the most saline water in this layer (salinities of more than 35.7 ‰) is found in the inner part of the ria. As we shall see this water can be regarded as resulting from an influx of ocean water along the bottom of the ria, ascending to smaller depths in the inner part. The bottom salinities observed in July 1968 by Gómez Gallego are generally somewhat lower (maximum value 35.64 ‰) than those discussed here.

In order to give a further idea of the vertical structure of the water masses of the ria, in fig. 6.4 the average vertical distribution is shown of temperature, salinity, density, and oxygen content of three groups of stations, called A, B and C. These groups are situated at the entrance, in the central part ('Rua deep') and in the inner part of the ria. These averages have been computed from 4,15 and 7 stations respectively. As may be seen, the properties of the water at the bottom are nearly the same in the three regions, although the depth decreases from about 70 m to about 20 m. This also illustrates the upward movement of water in the inner part of the ria, mentioned earlier in this section.

At the surface there is, as may be expected, a lower salinity in the inner part of the ria than at the entrance. The result is an increase of the vertical stability of the water-layers towards the inner parts. The variation of dissolved oxygen will be discussed in a later chapter, but is shown here together with the variation of the other properties.

6.2 Variability of the properties of the water

The different factors which influence the circulation of the water in the Ría de Arosa and which determine the physical and chemical properties of that water are not constant in time. As those factors are changing, there is also the possibility of a variation of these properties at a certain point. A presentation of the average situation is not sufficient as a description of the oceanographic conditions, as the variability is of equal importance.

As an example of this variability the frequency distributions of surface temperature and surface salinity of the water near the landing pier of the base camp at Punta Preguntoiro are given in Table VII and figs. 6.5 and 6.6. Although it can be assumed that the properties at this site are not fully representative for large parts of the ria outside the direct influence of the shore, and that they are more variable because of the small depth and certain small scale circulation patterns, these data still demonstrate the range of temperature and salinity that may be expected at comparable places along the shoreline, and may therefore be of interest from the ecological point of view.

On the other hand, places that can be expected to have a small range of temperature and salinity have to be looked for in the central parts of the ria at greater depths. Here the water is less affected by external influences. The variations that may occur at such places are demonstrated by the results of the observations made repeatedly in the 'Rua deep', in the central part of the ria between the islands of Rua and

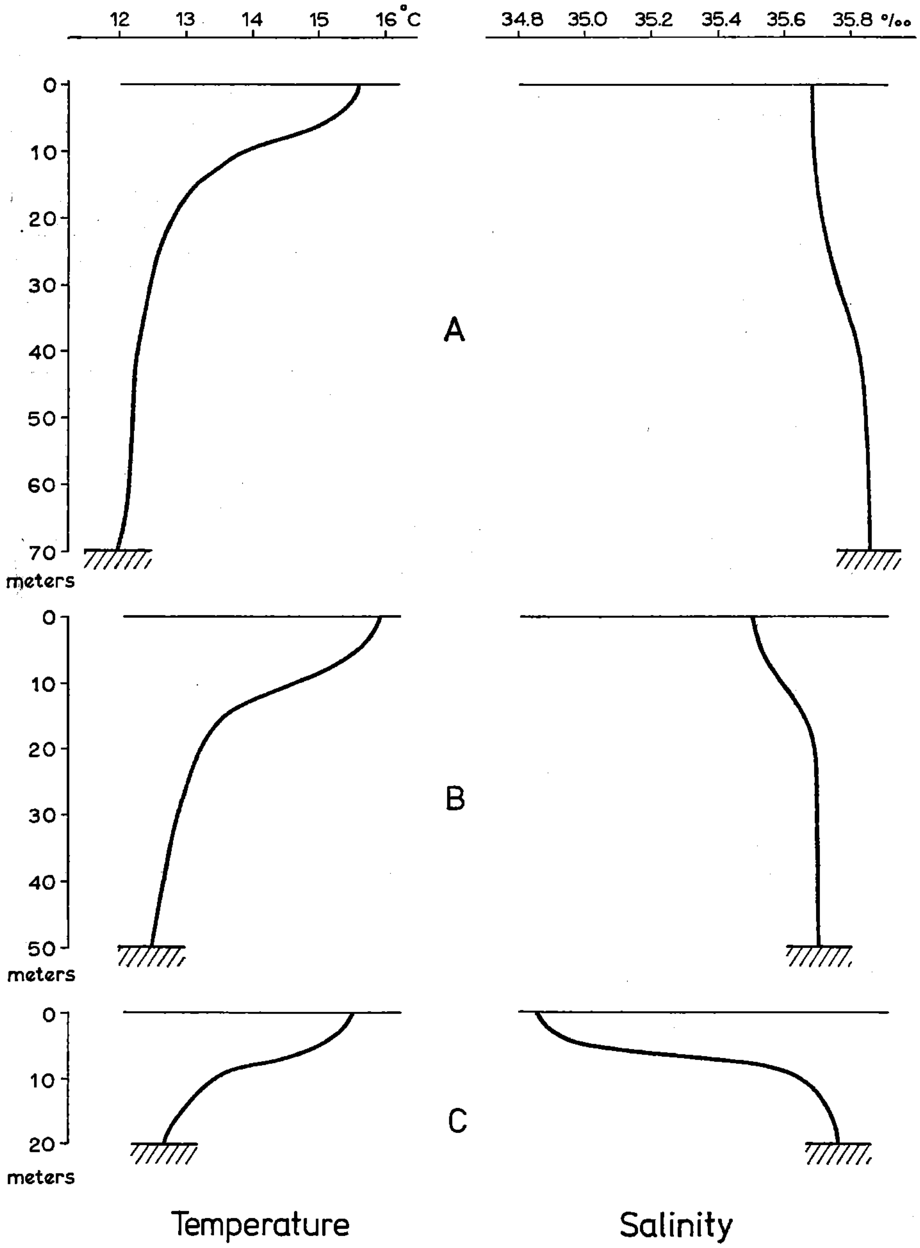
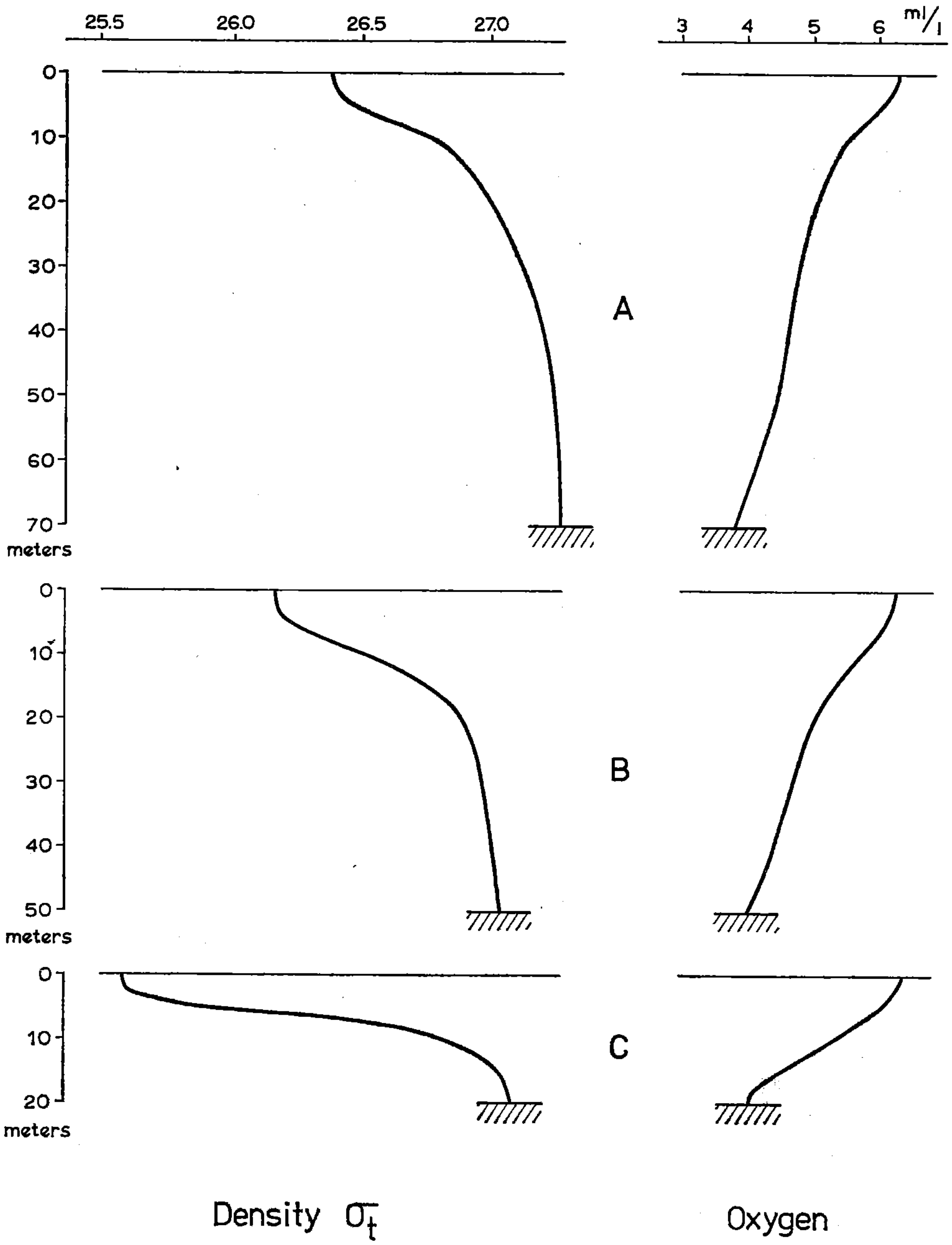


Fig. 6.4 Average vertical distribution of temperature, salinity, density and oxygen content at three points in the Ría de Arosa (summer).


25.5
26.0
26.5
27.0
3
4
5
6
ml/l
0
10
20
30
40
50
60
70
meters
A
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meters
B
0
10
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meters
C
Density σ_t
Oxygen

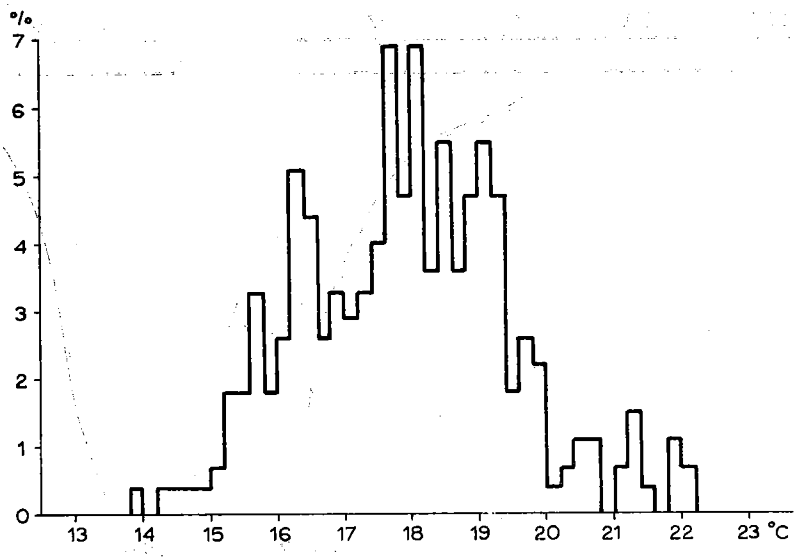


Fig. 6.5 Frequency distribution of surface temperature at Punta Preguntoiro (summer).

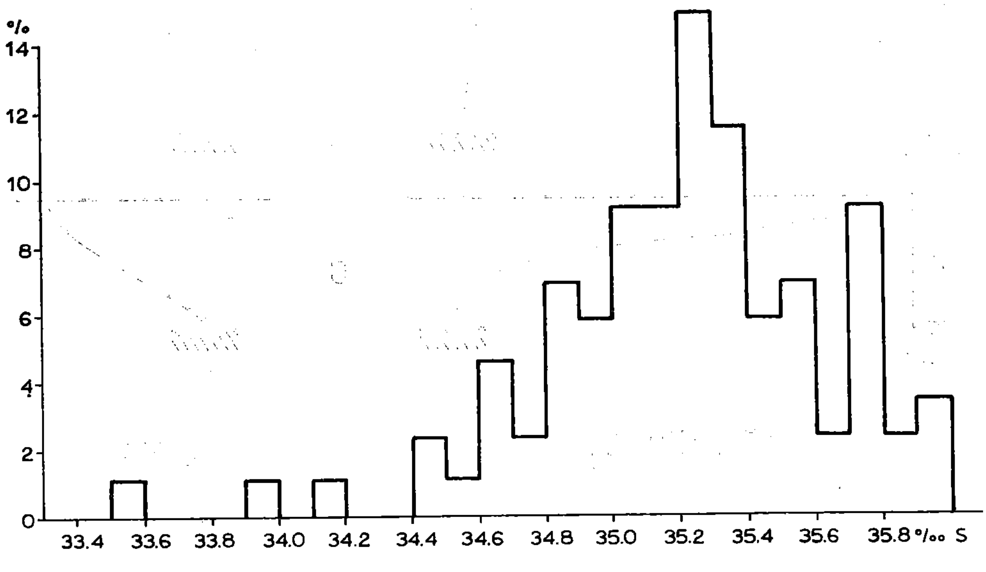


Fig. 6.6 Frequency distribution of surface salinity at Punta Preguntoiro (summer).

Jidoiro Pedegroso. The ranges of temperatures and salinities below 40 m are given in table VII.

TABLE VII.

Frequency distribution of surface temperatures at Punta Preguntoiro (274 obs.)

	.00-.19	.20-.39	.40-.59	.60-.79	.80-.99
13°	—	—	—	—	0.4
14°	—	0.4	0.4	0.4	0.4
15°	0.7	1.8	1.8	3.3	1.8
16°	2.6	5.1	4.4	2.6	3.3
17°	2.9	3.3	4.0	6.9	4.7
18°	6.9	3.6	5.5	3.6	4.7
19°	5.5	4.7	1.8	2.6	2.2
20°	0.4	0.7	1.1	1.1	—
21°	0.7	1.5	0.4	—	1.1
22°	0.7	—	—	—	—

Frequency distribution of surface salinities at Punta Preguntoiro (87 obs.)

	.00-.09	.10-.19	.20-.29	.30-.39	.40-.49	.50-.59	.60-.69	.70-.79	.80-.89	.90-.99
33⁰/oo	—	—	—	—	—	1.1	—	—	—	1.1
34⁰/oo	—	1.1	—	—	2.3	1.1	4.6	2.3	6.9	5.8
35⁰/oo	9.2	9.2	14.9	11.5	5.8	6.9	2.3	9.2	2.3	3.4

Frequency distributions of temperatures below 40 m between Rua and Jidoiro Pedegroso (16 obs.)

	.00-.19	.20-.39	.40-.59	.60-.79	.80-.99
11°	—	—	—	—	6
12°	19	12	31	25	—
13°	6	—	—	—	—

Frequency distribution of salinities below 40 m between Rua and Jidoiro Pedegroso (16 obs.)

	.00-.09	.10-.19	.20-.29	.30-.39	.40-.49	.50-.59	.60-.69	.70-.79	.80-.89	.90-.99
35⁰/oo	—	—	—	—	12	12	19	44	6	6

6.3 Characteristics of the water masses of the Ría de Arosa

Further insight in the properties of the water in the ria, their distribution, origin and circulation may be obtained from the study of the temperature-salinity (t-S) diagrams of the water at different levels. The observations of the three summer campaigns have been combined in diagrams for the layers of less than 10 m deep, 10-20 m, 20-40 m and more than 40 m deep. The results are shown in fig. 6.7 to 6.10.

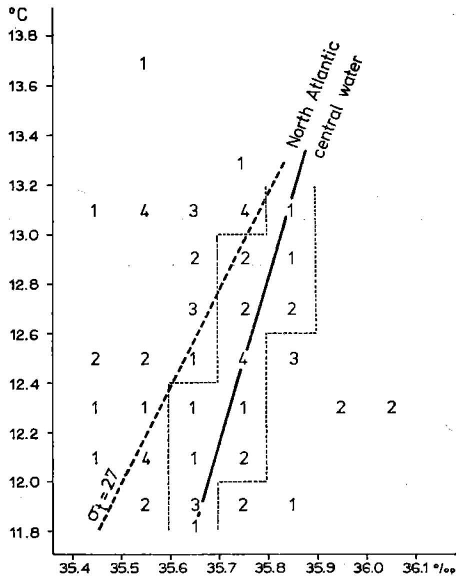


Fig. 6.7 *t-S* diagram for observations in the > 40 m layer of the Ría de Arosa (summer).

The regular observations made at Punta Preguntoiro, the frequency distribution of which is given in the previous section, are not included in these diagrams. In the figures the line characterizing the North Atlantic Central Water is also shown (see section 4.2). This water mass appears to occupy an important part of the waters of the Ría de Arosa, as follows from an inspection of the *t-S* diagrams.

In fig. 6.7 showing observations made at depths below 40 m, we see that 20 of them that is about 30 % of all observations, are arranged along the line for the N.A.C. water. Ten observations are found to the right of this line, with higher salinities than N.A.C. water of the same temperature. About half the number of observations are from warmer and somewhat less saline water than the N.A.C. water. As the observations are quite evenly distributed over the whole area having depths over 40 m, we may conclude that this distribution of observations reflects the volumetric distribution and that on the average about 30 % of the water volume below this depth consits of unmodified N.A.C. water.

The small number of observations with higher salinities indicates the presence of water from a different source of deep water. The temperature of this water is relatively low, and it is only found at sub-surface levels. Therefore this water cannot be surface water that acquired its high salinity by evaporation. We mention this possibility, because according to SAIZ ET AL. (1961) it sometimes plays a part in the shallow Ensenada de San Simón of the Ría de Vigo, and therefore at first sight might be considered a possible cause for high salinities in the other rias as well. A more probable source appears to be the high-salinity water, mentioned in section 4.5, possibly mixed with N.A.C. water. This water mass appears irregularly along the Spanish coast and is thought to originate further to the west.

The diagrams for the levels of 20-40 m and 10-20 m deep also show some contribution of N.A.C. water and incidentally also of the more saline water mentioned above, but they demonstrate especially how the water at these depths for a large part has much higher temperatures without an important change of salinity.

The diagram of the 0-10 m layer also shows this increase of temperature as compared with water at greater depths, and a great number of observations is found with

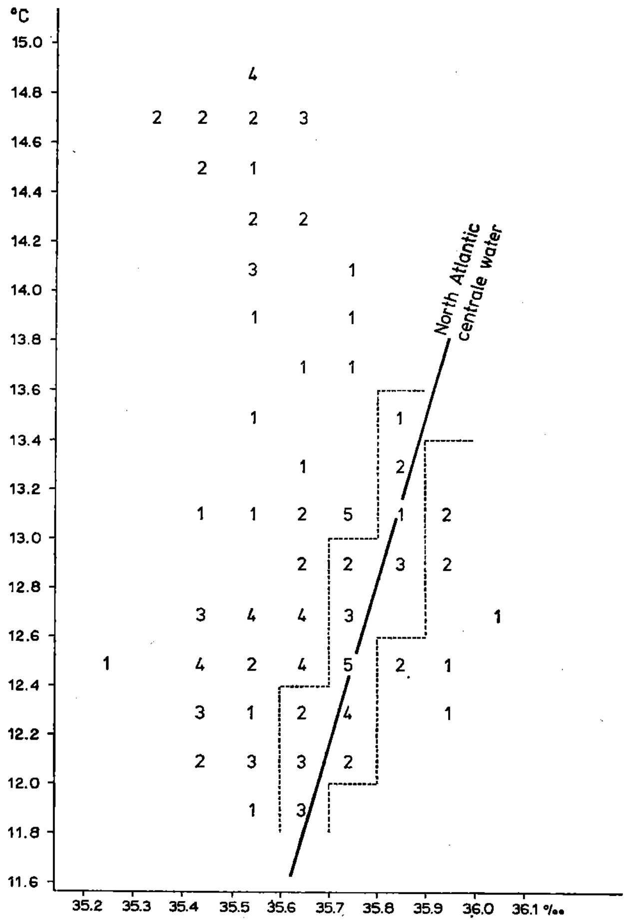


Fig. 6.8 *t-S* diagram for observations in the 20-40 m layer of the Ría de Arosa (summer).

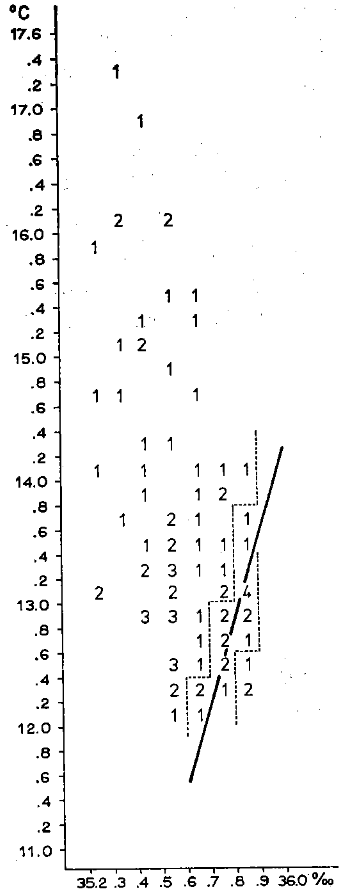


Fig. 6.9 *t-S* diagram for observations in the 10-20 m layer of the Ría de Arosa (summer).

temperatures between 16.0° and 17.0° and salinities between 35.2 ‰ and 35.6 ‰. In this area of the *t-S* diagram 65 observations are indicated, that is 20 % of all observations at those depths.

Further warming of this water is usually accompanied by some dilution with freshwater.

We here classify all observations with salinities between 34.8 ‰ and 35.6 ‰ and temperatures between 16.0° and 18.8° (45 % of all surface observations) as 'ria surface water'. Only a small number of observations is found with still higher temperatures and with not too low salinities (over 34 ‰); most of the other observations show an increased admixture of river water without a further increase in temperature.

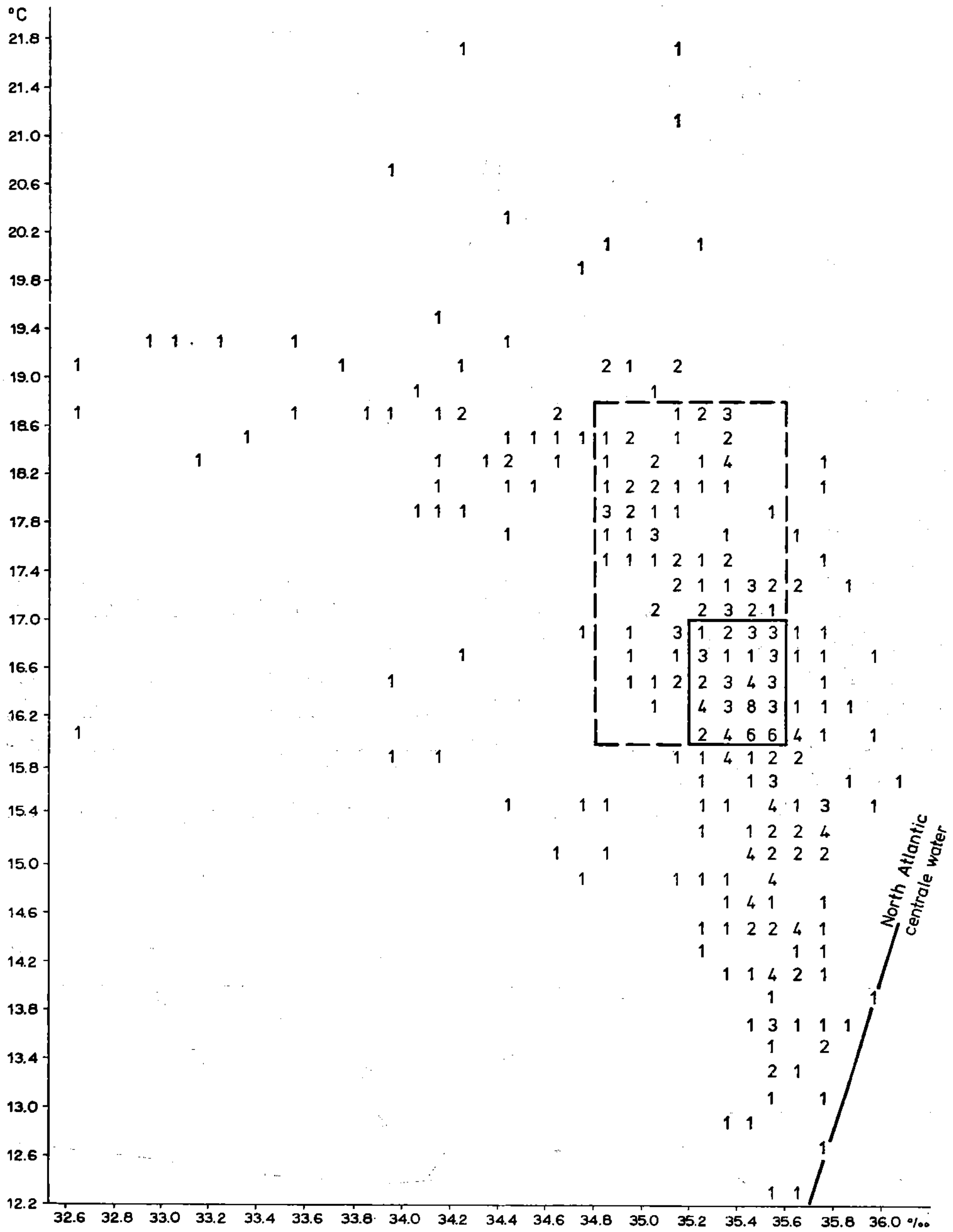


Fig. 6.10 *t-S* diagram for observations in the 0-10 m layer of the Ría de Arosa (summer).

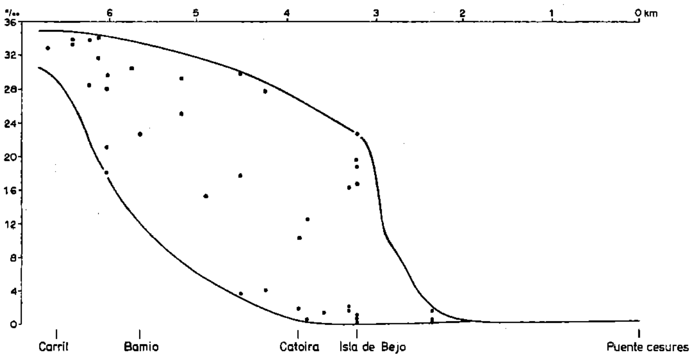


Fig. 6.11 Salinity distribution along the Rio Ulla (summer).

6.4 Properties of the river water

In the previous sections the influence of the rivers on the properties of the water has already been indicated. In the following the properties of the water in the rivers themselves (or at least in their brackish parts) will be studied.

The effect of the tide on the properties of the water at a certain point in the lower part of a river is very prominent. This is in contrast with the situation in the ria itself, where such an effect of the tide, although still of importance, often is largely counterbalanced by the influence of the wind, as will be shown later on.

The amounts of water which pass through a section of the lower part of the river

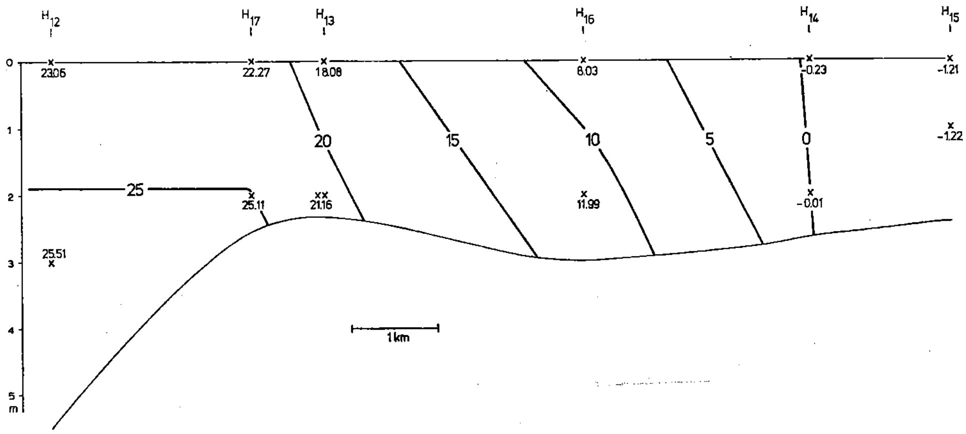


Fig. 6.12 σ_t profile along the Rio Ulla (3 July 1963).

during each tidal period exceed many times the river discharge. For instance, at Carril the tidal volume that flows up and down the river is estimated (from the surface area of the river upstream and the difference between high and low water along the river) to be about ten times the discharge in the summer. This means that a large variation of the salinity at this section may be expected during a tidal period as long as there is an important salinity gradient along the river.

In fig. 6.11 the salinities that were observed along the Rio Ulla during the summer campaigns at different phases of the tidal cycle were plotted, and the envelopes of this group of points were drawn, roughly indicating the variation of salinity that may be expected at the different points of the river. Of course some influence of variations in the run-off of the river gives an additional variation but as the measured values of the run-off differ by only 20 % this additional variation was thought to be of secondary importance compared with the variation caused by the tide.

The water in the river is slightly stratified, as is shown in fig. 6.12.

In fig. 6.13 a picture of the variation of the salinity along the Rio Uma is given.

The following temperatures were measured in the fresh river water of the Rio Ulla.

3 July 1963	18.6°
16 July 1962	22.5°
25 Aug. 1964	25.1°

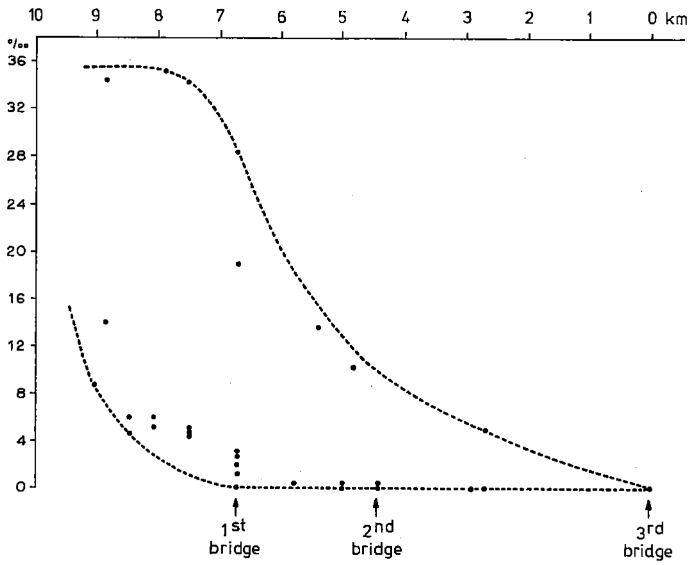


Fig. 6.13 Salinity distribution along the Rio Umia (summer).

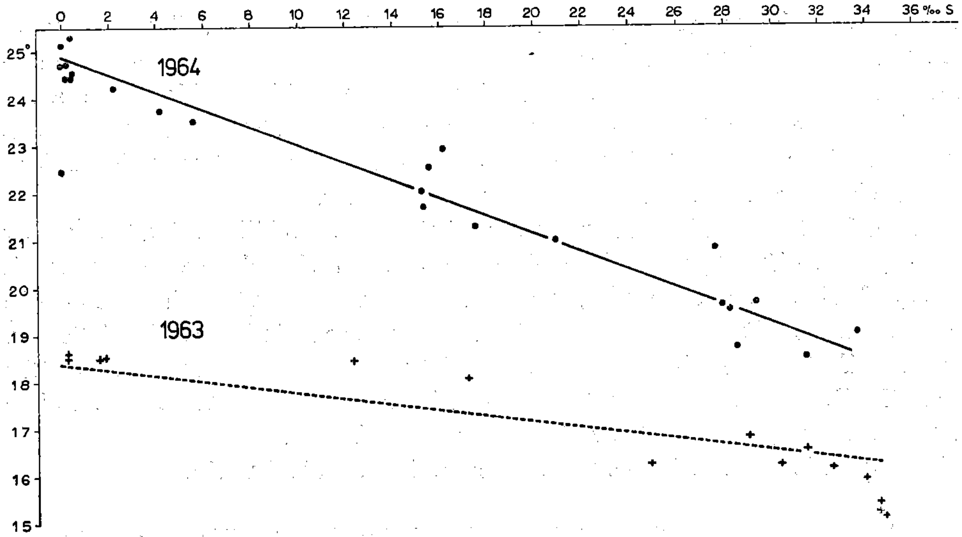


Fig. 6.14 *t-S* relation for Rio Ulla (summer).

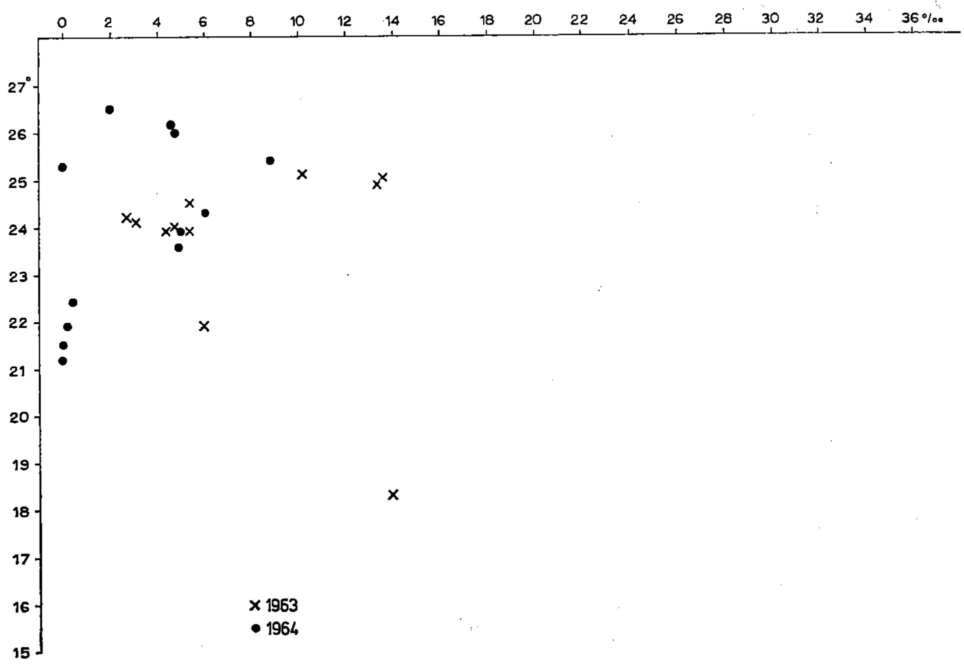


Fig. 6.15 *t-S* relation for Rio Umia (summer).

The temperature differences appear to be the result of a net advection of heat in the course of the summer.

In fig. 6.14 the temperature-salinity relation for the Rio Ulla is given, as found from two series of observations, on 3 July 1963 and on 5 August 1964. The points in this diagram are roughly arranged along straight lines, which indicates that the temperatures in the estuarine part of the river are primarily the result of the mixing of fresh water and sea water with different temperatures.

In fig. 6.15 a similar diagram is given for two series of observations on the Rio Umia, made on 9 July 1963 and on 29 July 1964. In contrast to the diagram of the Rio Ulla the points here show a much greater scattering, and it is not possible to find a well-fitting linear relation between temperature and salinity. This seems to be an indication that in the Rio Umia the temperature of the water is much more influenced by local heating and cooling. This may be ascribed to the small average depth of this river, where at low tide large banks and extended regions with shallow depths are found in the area where fresh and salt water mix. This results in a stronger heat exchange with the atmosphere than in the Rio Ulla with its larger depths and greater discharge.

6.5 Water of oceanic origin

As has been shown, the water flowing into the ria out of the ocean at sub-surface depths at least partly consists of North Atlantic Central Water. This water has a σ_t value just over 27.0 (see fig. 6.7). In the following we will therefore consider the $\sigma_t = 27.0$ isopycnic surface as the boundary between the water masses that belong to the North Atlantic Central Water or to what has been called the high-salinity water on the one hand, and the waters of a mixed type on the other hand. The average position of this surface, computed from our observations over the three summer periods, is given in fig. 6.16. The ascending of this surface in the inner part of the ria is striking. While in the outer parts this surface lies at depths between 30 and 40 m, it is found at less than 10 m from the surface in the inner parts of the ria.

It is interesting also to consider in this connection the concentration of dissolved oxygen at the isopycnic surface $\sigma_t = 27.0$. As this surface, except in the inner part of the ria, lies below the so-called compensation depth (see section 8.3) which means that oxygen consumption surpasses oxygen production by photosynthesis, the concentration of dissolved oxygen will be the result of an equilibrium between net oxygen consumption and advection of water with a higher oxygen content by mixing or by advective processes.

In fig. 6.17 the average oxygen content of the water at the $\sigma_t = 27.0$ surface is given. There is a gradual decrease from the mouth of the ria inwards. This can be interpreted as the result of consumption of oxygen while the water moves approximately along

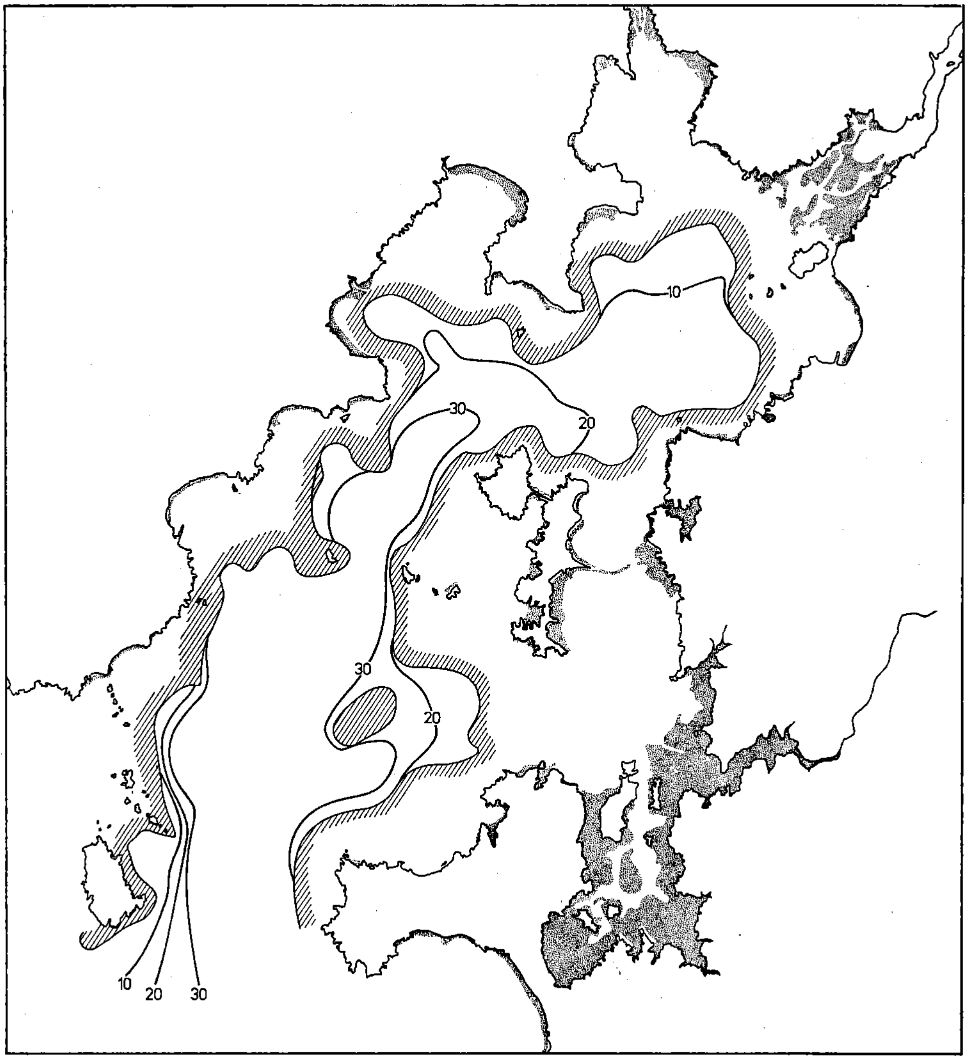


Fig. 6.16 Average depth of $\sigma_t = 27.0$ surface in the Ría de Arosa (summer).

the isopycnic surface. However, an exception is found in the eastern part of the inner ria. Here relatively high oxygen values are found. This may be caused by a reversal of net oxygen consumption to net oxygen production because of the small depth at which this particular surface is found here. We shall not dwell upon this problem here, the less so as the small number of observations makes an explanation at this moment speculative.

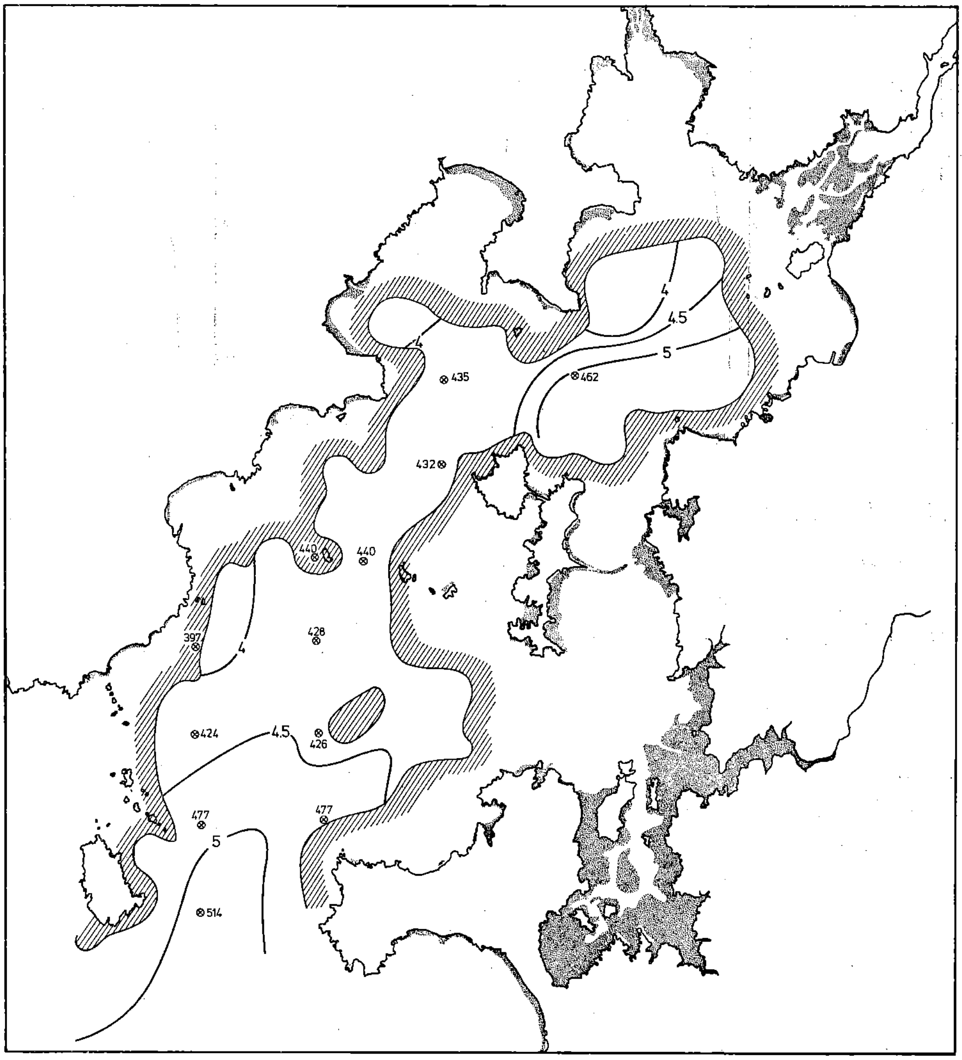


Fig. 6.17 Average oxygen concentration in the $\sigma_t = 27.0$ surface in the Ría de Arosa (summer).

Very low oxygen values have not been observed. The lowest concentrations found were 3.7 ml/l, but GÓMEZ GALLEGO (1971) mentions 3.2 ml/l as a minimum value observed.

Finally it should be pointed out that the discussion given here concerns the average situation. The actual situation at a given time may be different, and the presence of oceanic water certainly cannot always be demonstrated.

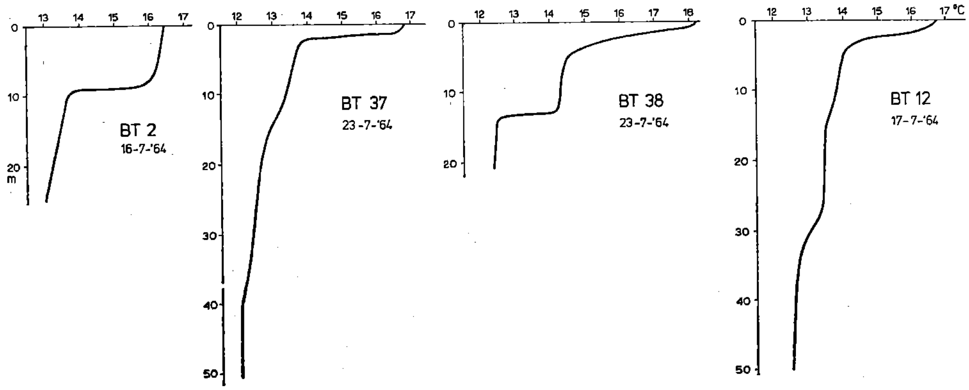


Fig. 6.18 Vertical temperature structure at some points in the Ría de Arosa.

6.6 Depth of the thermocline

The vertical variation of water temperature was investigated by means of bathythermograph observations giving greater detail than is possible with serial measurements. In fig. 6.18 some examples are given of the vertical structure of temperature. Often a thermocline at about 10 m could be observed, as is shown in the first example (BT 2). However, a very strong temperature gradient occurred also at much smaller depths, as the second example (BT 37) shows where the thermocline at 10 m depth is absent, and only a slight inflection of the curve is seen at about that depth. This strong, shallow temperature gradient may be ascribed to heating of the surface layers by incoming radiation or exchange with the atmosphere during very quiet weather. A combination of both thermoclines is seen in the third example (BT 38). The last example shows a situation that was sometimes observed, with an inflection of the temperature curve around 30 m.

In fig. 6.19 the frequency of occurrence of thermoclines or marked inflections of the temperature curve is shown for the inner, central and outer ria, respectively. Especially in the inner and central ria marked temperature gradients are frequently observed at two distinct levels.

One other point deserves attention. On the BT slides often a doubling of the trace was observed. Such a double trace may be caused by instrumental defects (hysteresis in the temperature or pressure recording) or by real short-time variations in the vertical temperature distribution. As at other occasions such an effect was not observed, and as lowering and hoistring was always done slowly, the observed effect is thought to be a real variation of temperature, caused by internal waves. Traces often showed a vertical displacement of isotherms of the order of 2 m. The time between lowering and hoisting usually was about 10-15 min. From the vertical variation of density we may calculate the period of standing internal transverse oscillations (internal seiches).

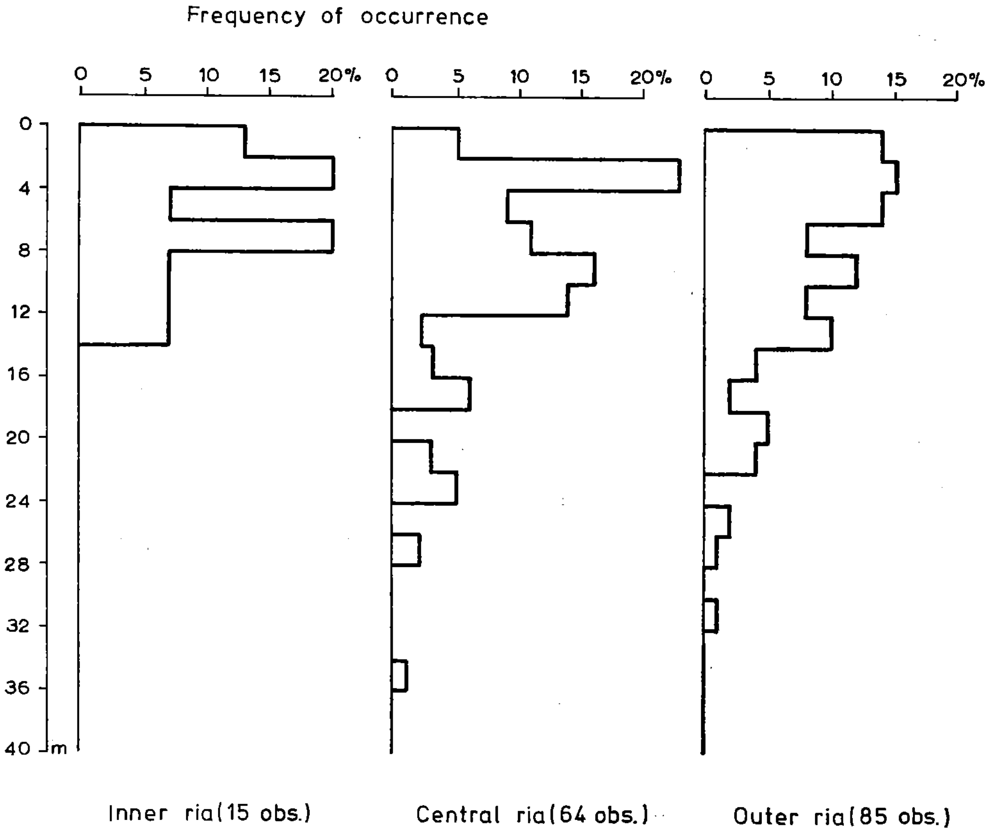


Fig. 6.19 Frequency of the occurrence of the thermocline at different depths.

The first mode of such oscillations for a two layered channel with depth d, width B and a density ρ_1 of the upper layer with thickness h and a density ρ_2 of the lower layer (thickness $d - h$) is approximately given by the relation

$$T = 2B\sqrt{\frac{\rho_2 d}{g(\rho_2 - \rho_1)\, h\, (d - h)}}$$

In the present case this gives a period of about 1-3 hours. This result is not incompatible with the BT observations.

6.7 Circulation of water in the Ría de Arosa

The results of the discussions in the foregoing paragraphs already suggest a certain type of circulation. The maps giving the average distribution of salinity and temperature show how the river water, gradually mixing with ria water, flows outward in the surface layers, mainly along the right shore. Furthermore we found an influx of water from the ocean, either permanent or intermittent, along the bottom, without much dilution with water from higher layers and possibly along the left side of the main channel.

A process of entrainment of sub-surface water by the surface layers may to some degree be enforced by the influence of the wind, when blowing from northern directions, as will be demonstrated in the next paragraphs. This may even go so far that the circulation process can better be considered to be an upwelling process than a process of estuarine circulation.

For such cases BRONGERSMA-SANDERS (1965) has used the term 'estuarinelike circulation'.

It may be mentioned that the effect of the wind on the circulation has also been reported for the Ría de Vigo by SAIZ ET AL. (1961) and by ANADÓN ET AL. (1961).

6.8 Some reports on currents in the Ría de Arosa

Data on the currents in the Ría de Arosa are scarce. A value of one knot (0.5 m/s) for the maximum velocity that can be found is given in the 'West Coast of Spain and Portugal Pilot' (HYDROGRAPHIC DEPARTMENT, 1957).

More specific information on currents is given by NONN (1966) who quotes an unpublished report of the director of the harbour of Villagarcia. According to these data the current between the peninsula of El Grove and Isla de Salvora attains values of about 35 cm/s, while east of the southern point of Isla de Arosa the current velocity would attain values of only about 15 cm/s.

Recently GÓMEZ GALLEGO (1971) has published current data from different depths for a number of stations in the ria. Current speeds up to 35 cm/s were observed. In a map this author shows the direction and speed of the current at different points during both flood and ebb.

In their study of the Ría de Vigo, SAIZ ET AL. (1957) have made some general remarks concerning the current in the Rías Bajas. They have used a simple model of the Galician west coast in order to study the tidal currents in this region. It is stated that the ebb current leaves the Ría de Arosa, analogous to the Ría de Pontevedra and the Ría de Vigo, through the southern entrance, that is between the island of Sálvora and the peninsula of El Grove. The flood current would not show such a preference. However, the authors do not give particulars on the model used, and presumably the Coriolis force is not taken into consideration. Therefore their data should be considered with caution.

6.9 Average velocity of the tidal currents

Data on the vertical tide in the Ría de Arosa are presented in the tidal tables published for instance by the DEUTSCHE HYDROGRAPHISCHE INSTITUT (see section 1.6.)

The average velocity of the inwards and outwards directed components of the tidal currents in a certain cross-section may in our case be calculated from the tidal amplitude, the dimensions of this cross-section and the dimensions of the surface area of the ria landwardof this cross-section, for the volume of water passing this section is equal to the tidal prism of the landward area.

If D is the cross-sectional area and if the part of the ria landward of the cross-section has a surface O, and if the vertical tide is approximated by a simple cosine function with period T and amplitude A, the average current velocity in the cross-section is given by

$$\bar{u} = \frac{2\pi O \cdot A}{T \cdot D} \sin 2\pi \frac{t}{T}$$

and the maximum value of $\bar{u}$ is given by

$$\bar{u}_{\max} = \frac{2\pi O \cdot A}{T \cdot D}$$

From calculations for different cross-sections of the ria it was found that for the main part of the ria the values of $\bar{u}_{\max}$ vary between about 5-10 cm/s during neap tide

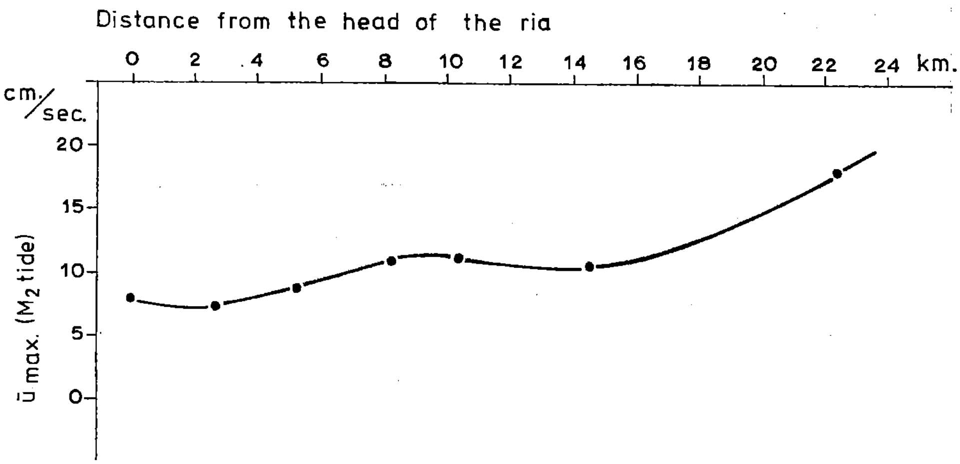


Fig. 6.20 Variation of the estimated mean longitudinal component of the tidal current along the axis of the Ría de Arosa.

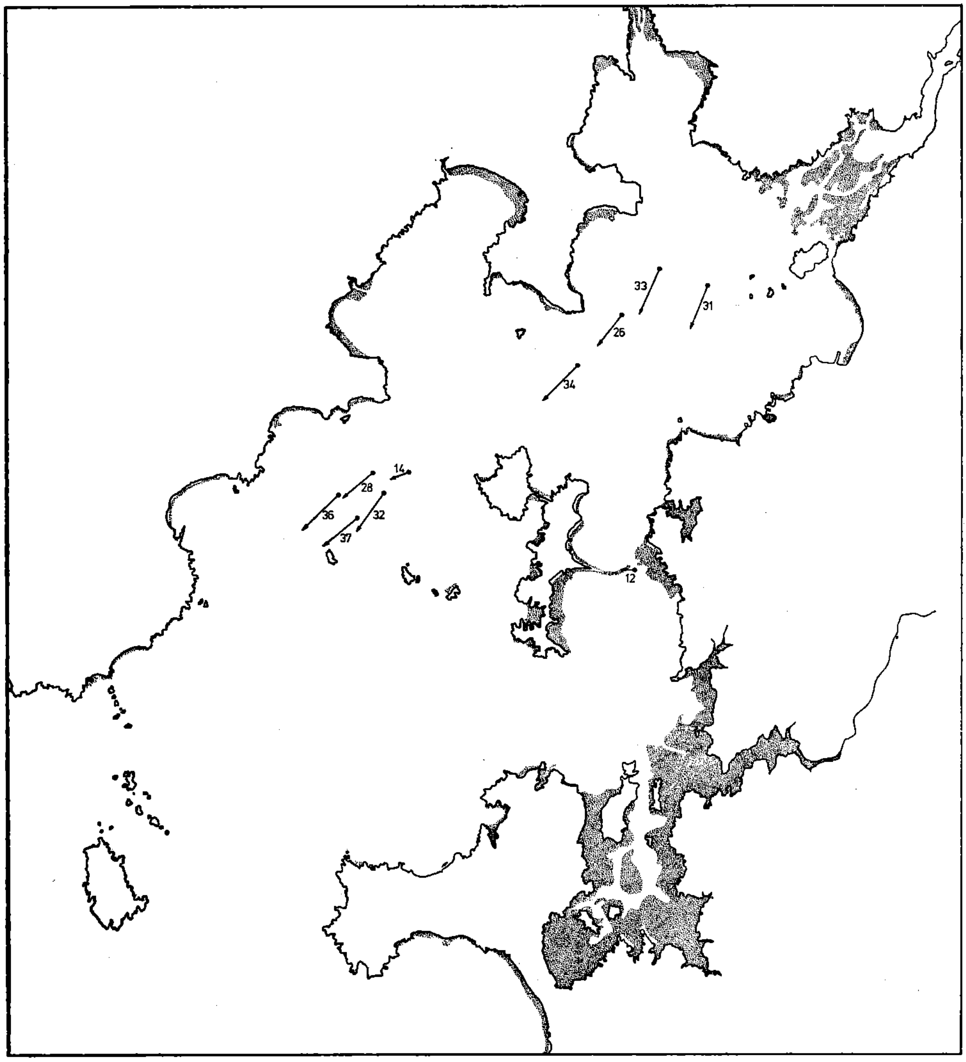


Fig. 6.21 Surface current in the Ría de Arosa, ebb, N-ly wind.

and 10-20 cm/s during spring tide. Only for the north-south directed cross-section between El Grove and Isla de Arosa values were found that are two times as high.

The variation of $\bar{u}_{max}$ under average tidal conditions (M_2 tide) along the length of the ria is given in fig. 6.20.

It should of course be kept in mind that these are average values of the current components perpendicular to the cross-sections. The actual currents in different parts

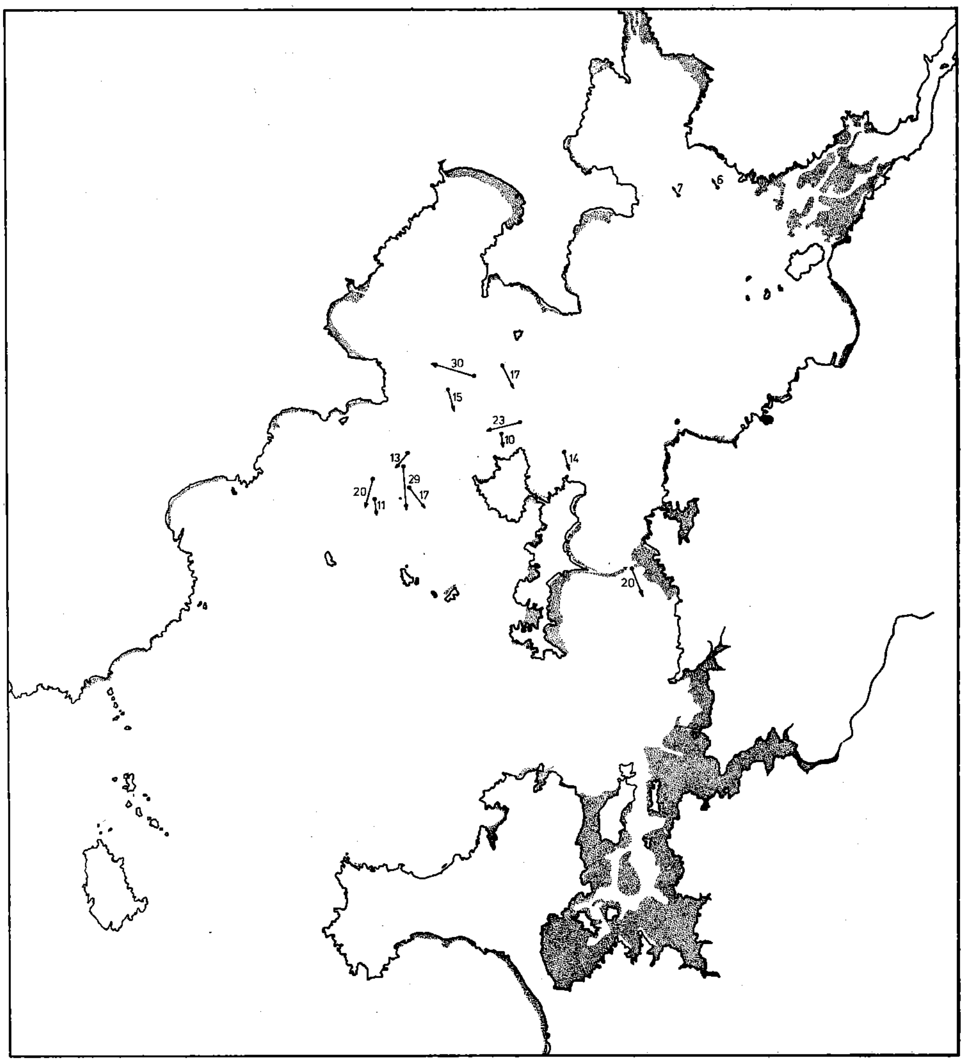


Fig. 6.22 Surface current in the Ría de Arosa, flood, N-ly wind.

may have different values, higher and lower than the calculated average, or even opposed to the direction of mean transport.

However, knowledge of these average values is useful for comparison with actual observations. If large deviations from the mean value are found, we may conclude that there are important vertical or lateral differences in the current.

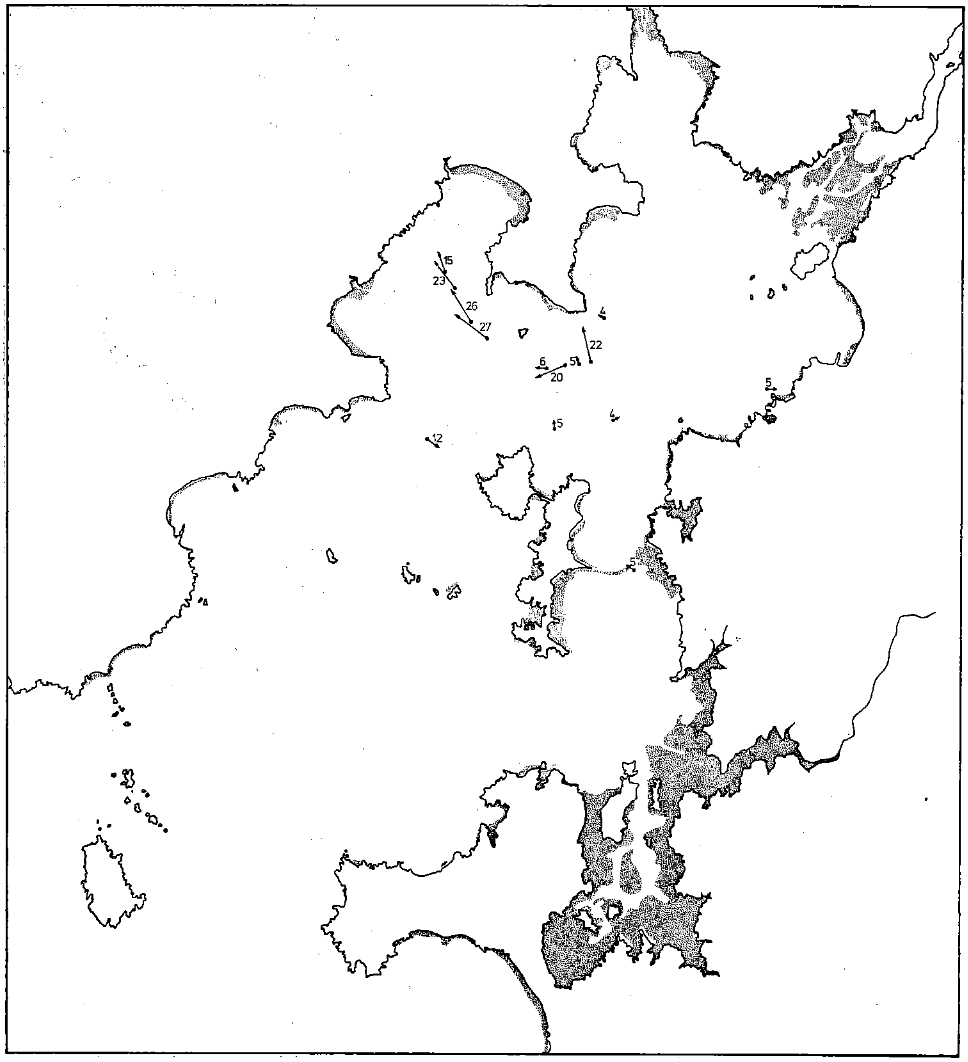


Fig. 6.23 Surface current in the Ría de Arosa, ebb, SW-ly wind.

6.10 Observed surface currents under different conditions of tide and wind

During the course of the summer campaigns surface current measurements were made at various occasions. These measurements were made with current meters from anchored vessels and with drifters. The results of these measurements were combined into four groups, according to the tidal phase (flood or ebb) and to the wind direction

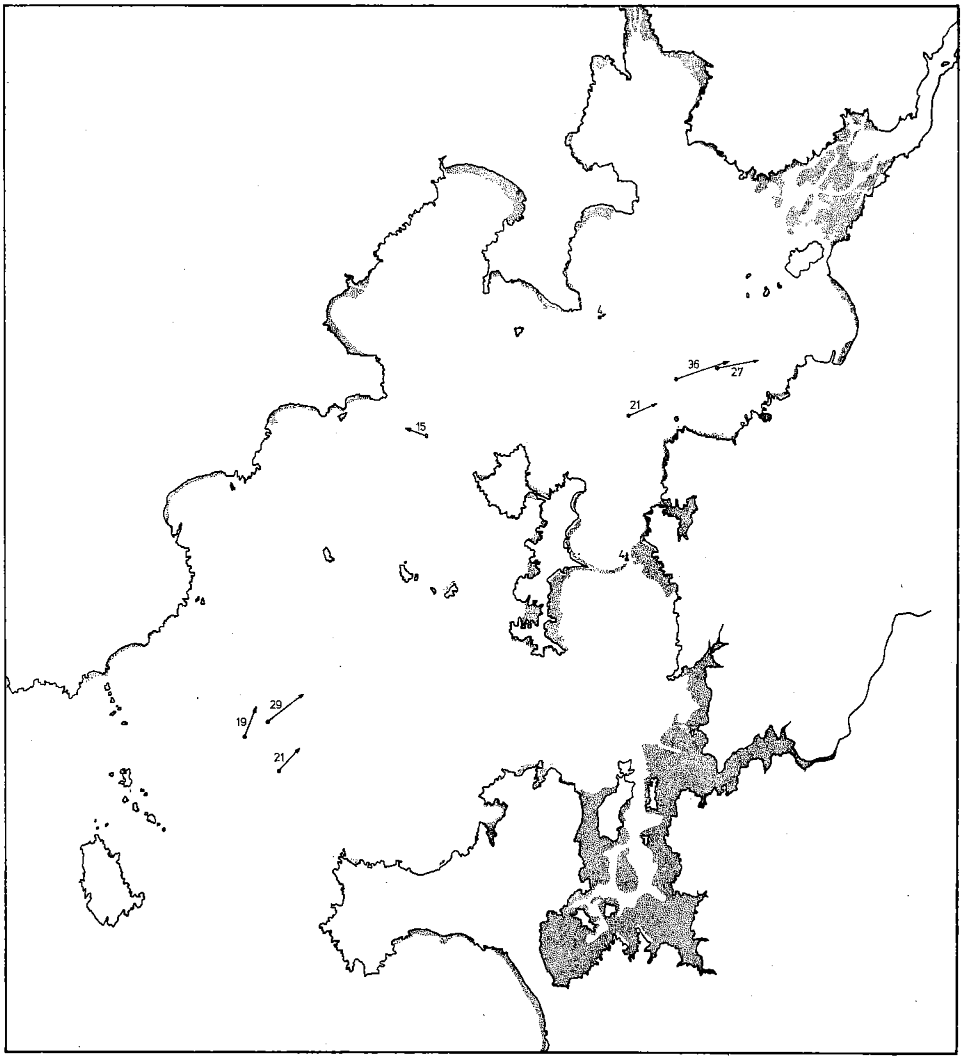


Fig. 6.24 Surface current in the Ría de Arosa, flood, SW-ly wind.

(northerly and southwesterly winds). In fig. 6.21-6.24 the currents in these four situations are shown. The following may be remarked in connection with these results.

1. Ebb current, northerly wind

The surface current in the main part of the ria flows outward. The observed velocities are on the average about 30 cm/s. This is more than the average value calculated for a whole cross-section. Therefore in other parts, either at the surface outside the main axis, or at greater depths the currents are appreciably less than the average.

The small westward current in the shallow part between the Isla de Arosa and the eastern shore near El Vado is not so easy to explain without additional observations.

2. Flood current, northerly wind

The current pattern appears to be very complicated with apparently various eddies. In the central part the current at the surface still has an outward component, apparently under the influence of the wind, with velocities of about 10-20 cm/s. This means that there should be a strong inward flow, probably at greater depths, with ascending motions in the inner parts of the ria to compensate for the outward movement.

At El Vado the northerly winds give rise to a moderate southerly current as might be expected in this shallow region.

3. Ebb current, southwesterly wind

Only a small number of current observations was made in the central part of the ria. However, they give a strong indication that the opposition between the wind direction and the tidal flow to a certain degree hampers the outflowing current, causing currents in transverse directions. In this case a major outflow may be expected closer to the shore or at greater depths.

The current near El Vado is very weak, which is also the case further north, west of Las Sinas. On the other hand there is a strong current towards the inner part of the bay of Puebla de Caramiñal, and a compensating current will probably exist along the southwestern shore or at greater depths.

4. Flood current, southwesterly wind

In the southern part the inflowing current is only slightly faster than the average for that part of the ria. In the central part there is only one observation.

The observed westerly current cannot easily been explained without additional observations. In the inner part the current appears to be concentrated along the southeastern shore, the current velocity in the northern part being negligibly small. At El Vado a weak northerly current is found.

Comparison of these data with those of Gómez Gallego is difficult as this author does not specify the wind conditions during his observations. His vector diagrams suggest a sometimes considerable variability. In the current map presented with the study of Gómez Gallego some noteworthy features appear.

1. East and northeast of Isla de Arosa the flood and ebb currents both have northerly components. This suggests the presence of a strong compensating net water transport north of Isla de Arosa, either by a large current eddy or by advection from larger depths.

2. The observation of an outward ebb current at the entrance of the bay of Puebla de Caramiñal indicates that the inward ebb current shown in fig. 6.23 is probably due to the southwesterly wind.

The component of the residual current at the surface in the direction of the main channel of the ria, i.e. the outflowing (here positive) current or the inflowing (negative) current may be estimated by subtracting from the component of the total current the tidal current estimated by the method described in paragraph 6.9. Application of this method presumes that vertical variation of the tidal current is small for the larger upper part of the water column and that the influence of lateral variations of the tidal current is more or less compensated by taking averages over a large number of observations made at different times and places. The mean residual surface current for the inner and central parts of the ria and the standard deviation of the individual estimates are:

moderate northerly winds	$+19$ cm/s ($\sigma = 9$ cm/s)
moderate southwesterly winds	$-\ 7$ cm/s ($\sigma = 9$ cm/s)

The rather large standard deviation is in part the result of the variations in windforce and of regional variation of residual and tidal current from place to place because of the irregular topography. The estimates should be considered indications of the order of magnitude only.

6.11 Currents at greater depths

The distribution of water properties and the results of current observations at the surface have already indicated that, at least under certain conditions, the current at sub-surface levels may deviate appreciably from the surface currents.

In the following section a number of deep current observations will be discussed that confirm this deduction. In the first place some observations made at different depths under conditions of variable and light winds will show that even without an important influence of the wind differences occur, presumably owing to friction and density distribution. These observations, made on 11 August 1964 in the wide outer part of the ria, southwest of the island of Rua, demonstrate the differences between the currents at 1 m and 30 m depth. They were carried out with two pairs of drifters simultaneously, each pair consisting of drifters with current crosses at both depths mentioned above. The two pairs were launched about 1 km apart. The results obtained during the first hour of the ebb period for both pairs show deviations to the right of the deeper currents of 32° and 25° respectively. The current velocities were 9 cm/s for both drifters of one pair and 15 cm/s and 6 cm/s for the surface and deep drifter of the other pair.

At higher wind forces we may expect additional effects of the wind, which may give rise to a circulation pattern with a surface and deep flow running in opposed directions In Table VIII some data are summarized that were obtained by measurements with pairs of drifters at different depths under various conditions of wind and tide. Although they only refer to situations with moderate winds they illustrate the differences

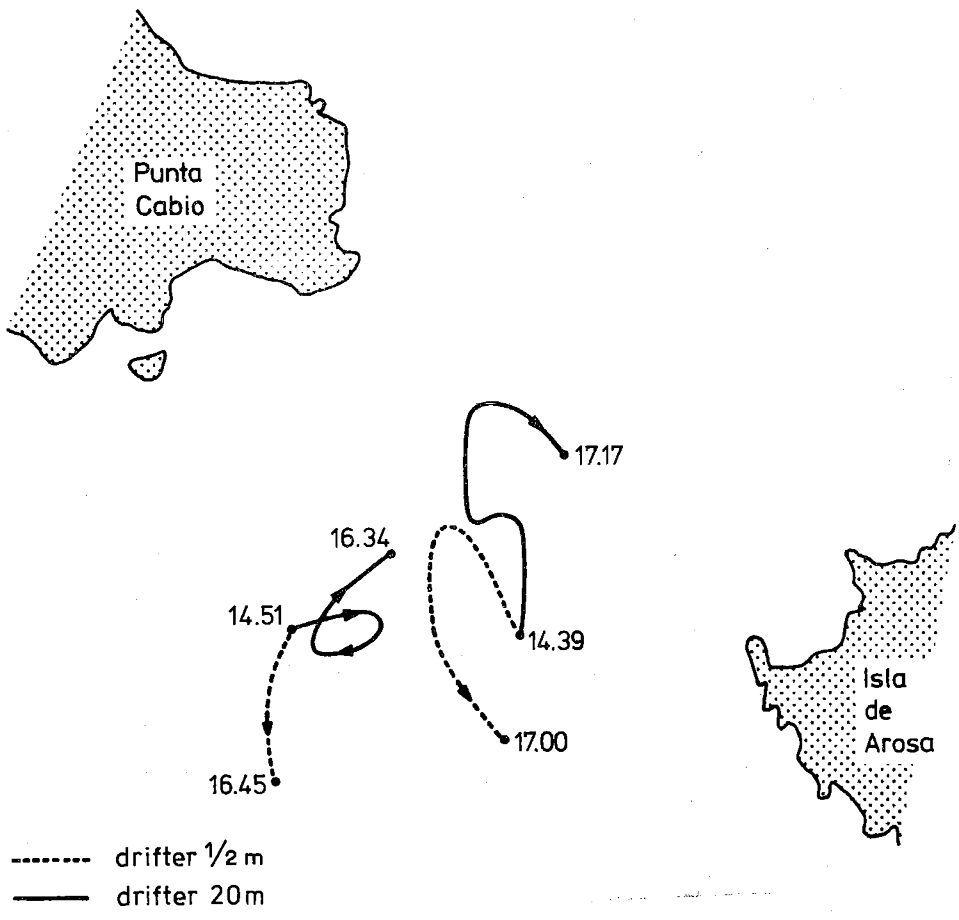


Fig. 6.25 Current observations, 30 July 1964.

that occur. The mean longitudinal components (SW and NE) of the observed currents are given and by subtracting estimates of the mean tidal current over the different periods of the observations (see the method in section 6.9), the longitudinal components of the residual current were estimated too. In the cases in which the wind was blowing in about the same direction as the tide is running, the differences in the directions of the surface and deep currents were small, but when the wind had an opposite direction, large differences in the current direction were observed (120°-180° for N-ly wind and flood, and about 90° for SW-ly wind and ebb). As an example the observations made on July 30, 1964 are shown in more detail in fig. 6.25. During the flood current with northerly wind the drifters followed a complicated path, the average current at the surface being southward, apparently under influence of the wind, while the current at 20 m depth was running towards the inner part of the ria under influence of the tide.

The measurements discussed here strictly speaking have only bearing on the horizontal component of the deep current. Vertical movements cannot be measured in this way. However, if we may assume that the drifters exactly follow the horizontal component of the water movement and that the effect of the surface current and the wind on the surface float may be neglected, we may discover simultaneous upward or downward movements of the water in the following way.

As near as possible to the drifter temperature and salinity observations were made at the same depth as the depth to which the drifter reaches. If the water density between two such observations remains the same, vertical movements of the water, if existing, will be small. If the density decreases or increases, this is an indication of a downward or upward movement of the water, respectively. If mixing may be neglected and the current is along the isopycnic surfaces, the variation in the density and the vertical density gradient give an estimate of the vertical velocity.

TABLE VIII. *Data on simultaneous current measurements at different depths, made under various conditions of wind and tide*

day	region	wind	tide	mean bottom depth	surf. current (cm/s)	residual current (cm/s)	depth deep drifter	deep current (cm/s)	deviation direction current	with surf. residual current (cm/s)
30 July 1964	Centr. Ria	N, 4Bft	ebb	30-60 m	33-27	+17	20 m	15-7	small	—0.5
		N, 4Bft	flood	30-60 m	12-7		20 m	12-11	120°-180°	
13 Aug. 1964	Inner Ria	SW, 2-3Bft	ebb	appr. 20 m	5-7	—10	10 m	3-6	about 90°	—2.5
		SW, 2-3Bft	flood	appr. 20 m	22-30		10 m	13-16	small	

On one occasion an attempt was made to investigate the presence of vertical movements in this way. On 7 August 1964 during flood, with a wind that increased from nearly zero velocity to Bft. 4 (northwest), temperature and salinity observations were made at the start and end points of two drifters at 30 m depth. For one drifter at the end point no observation could be obtained at the same depth as that of the drifter, owing to irregular bottom topography.

The results of the density determinations for the other drifter are:

start: 10.50 h, $\sigma_t = 26.94$
end: 13.36 h, $\sigma_t = 27.02$

horizontal drift between 10.55 and 13.24 h: velocity 16 cm/s, direction varying between 22° and 56°.

Thus, if the suppositions given above are acceptable, we may conclude that there was an upward movement of the water. As the average gradient of the density in the same density interval calculated from observations made at the same and preceding days in the same area gives an average increase of σ_t by 0.005 for an increase of depth of 1 m, (with a maximum increase of 0.007 and a minimum of 0.004), the upward movement is estimated to be

$$\frac{27.02 - 26.94}{0.005}\text{m, or 16 m in 2h 46 min. or 6m/hour or 0.16 cm/s.}$$

The results for the first-mentioned drifter are:

start: 11h 27 min. $\sigma_t = 27.08$ (at 30 m)
end: 14h 38 min. $\sigma_t = 27.08$ (at 30 m, estimated from the observation at 21 m and the vertical density gradient)

horizontal drift between 11h 29 min. and 14h 29 min.: mean velocity 14.5 cm/s, direction varying between 42° and 67°.

In this case there is therefore no indication of a vertical movement of the water.

The first value (0.16 cm/s) appears to be rather high, if we compare it with the figures that are deduced later on in section 10.6 (table XVII). However, those figures give estimates of the vertical movement over large areas and over a long period of time. Local and temporal vertical movements of higher velocity may occur.

The data presented in section 6.14 show vertical variations of σ_t surfaces that, if caused by internal waves, would be the result of vertical velocities of the same order of magnitude.

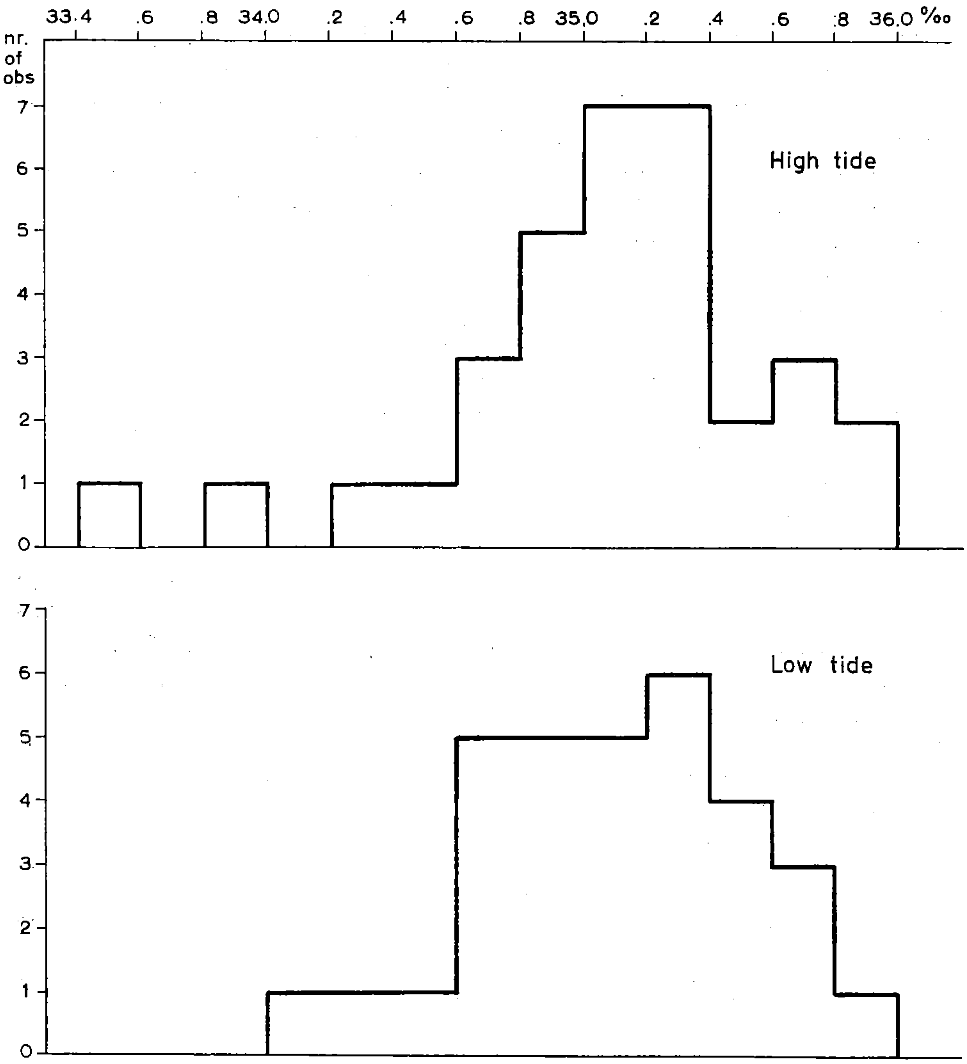


Fig. 6.26 Frequency distribution of salinity during high and low tide (Punta Preguntoiro, summer)

6.12 Variation of the water properties with variations of tide and wind

As both tide and wind influence the circulation of the water, it may be expected that these influences will be reflected in the distribution and local variation of the properties of the water. However, it is not certain whether such variations can be detected in rather small series of observations, because of other, more or less irregular, variations.

For a long series of observations made at one point, such as those made at Punta Preguntoiro, the average value of the salinity during different phases of the tide can be determined. If we then may assume that the effect of the wind will be largely cancelled out in these averages, we may attempt to determine the effect of the tide.

In fig. 6.26 the variation of the individual observations at high tide and at low tide is given. It is apparent that a tidal variation of the salinity, if existing, is completely obscured by other variations.

Near Punta del Chazo, on the northwestern side of the inner part of the ria, opposite Punta Preguntoiro, a relatively large number of observations was made as well. Although larger variations of the salinity may be expected (and indeed do exist) owing to the situation of this point in the axis of the river outflow (see fig. 6.1), the average tidal variation of the salinity could not be shown to be significant.

However, on individual days a definite tidal variation of the order of 1 $^0/_{00}$ was observed in this area by GÓMEZ GALLEGO (1971) with low and high salinities coinciding more or less with the times of low and high tide respectively.

The influence of the wind on the variation of the salinity at a point such as Punta Preguntoiro may be caused by different factors. The horizontal advection by wind-induced currents, the vertical advection by upwelling of deeper water and the increased vertical mixing at higher wind forces may all contribute to a wind effect.

The following table gives the average values of the salinity at Punta Preguntoiro under different wind conditions.

Wind	S-SW > 2 Bft	0 and 1 Bft	NW-NE > 1 Bft	NW-NE > 4 Bft
mean salinity:	35.14 $^0/_{00}$	35.20 $^0/_{00}$	35.31 $^0/_{00}$	35.41 $^0/_{00}$
stand. dev.:	0.37 $^0/_{00}$	0.50 $^0/_{00}$	0.55 $^0/_{00}$	0.35 $^0/_{00}$
number of observ.	22	8	29	7

The average values of the salinity are higher at northerly winds, as compared with southwesterly winds and calms, as might be anticipated if the upwelling caused by the northerly winds has an important influence on the conditions at this location. However, the salinity values show a large spreading and the increase of salinity is not significant.

Although in this section we did not find unambiguous indications of the effect of tide and wind on the properties of the water at the locations where longer series of observations were made, it does not mean that these effects do not exist; it has been shown before that the circulation is influenced by tide and wind to a considerable extent. The failure to demonstrate these effects on the water properties may be due to various causes. It may be that the different effects of, for instance, the wind to a large extent compensate each other, it may be that the variations or the number of observations are too small to give significant results because of other variations of an irregular character. In the following sections the variations that may occur will be further studied using observations made during half a day or a whole day at anchor stations, or made at stations repeatedly occupied in the course of a tidal period.

6.13 Observations made at anchorstations

An anchor station was occupied in the central part of the ria in July 11, 1962. The observations were made during a whole tidal period. The wind increased from calm at the beginning of the observations to about 5 Bft from westerly and southwesterly directions during the late afternoon. The variations of salinity, temperature, dissolved oxygen and density are shown in fig. 6.27.

It is seen from these data that at all depths the salinity exhibits rather large variations that cannot unambiguously be connected with either tidal or wind effects. The variations of the temperature are more regular. The highest temperatures of the surface water occur durong low tide. This does not need to be a tidal effect, the diurnal variation of the water temperature will have an influence too.

At greater depths the variations of the water temperature are very small, the values ranging from 13.08° to 12.98 °C, except one, perhaps erroneous, value of 13.21°.

The variations in the deeper water may be better studied from the data on dissolved oxygen content and density. Water with low oxygen values is present around the time of low tide. This water probably comes from the inner parts of the ria, as there is generally a decrease of dissolved oxygen from the mouth of the ria inwards. Water with $\sigma_t = 27.0$ (previously defined as ocean water) was incidentally observed, preferentially around the time of low tide. On this occasion this oceanic water probably had a limited extension in the ria, perhaps in the form of a more or less isolated patch at the bottom of the inner ria; the oxygen content and the presumed circulation of deep water under these conditions of tide and wind point in this direction.

Another set of data of interest is the group of observations made in september 1950 on an anchor station in the mouth of the ria by the Spanish research vessel 'Xauen' (INSTITUTO ESPAÑOL DE OCEANOGRAFIA, 1955).

These observations have been made in a different month and, as the situation in 1950 may have been quite different from the one in the years 1962-1964, these data are not directly comparable with our findings. Nevertheless they are very valuable, as they were made at two successive tidal periods and so give an impression of not only tidal, but also of diurnal variations (see fig. 6.28).

As regards the surface salinity we see the minimum values appearing during flood, while the maximum salinities are observed at low tide or during the ebb period. The temperature at the surface shows a maximum at both low tides, at 23.00 h and at 11.00 h, respectively, which clearly demonstrates that this effect is caused by advection (presumably from the inner part of the ria) and not by diurnal warming of the water by the sun or by exchange with the atmosphere. The slight increase of the

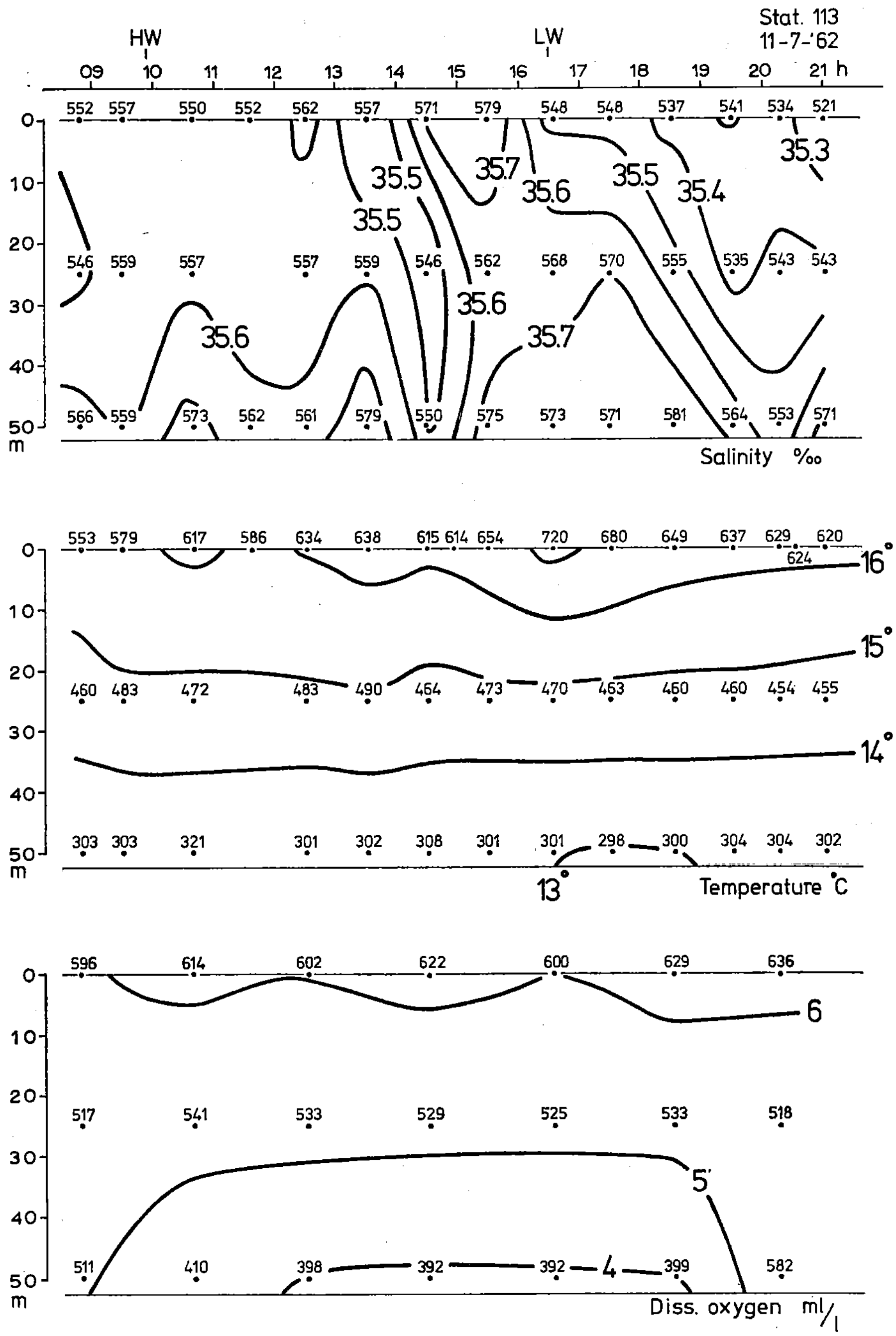
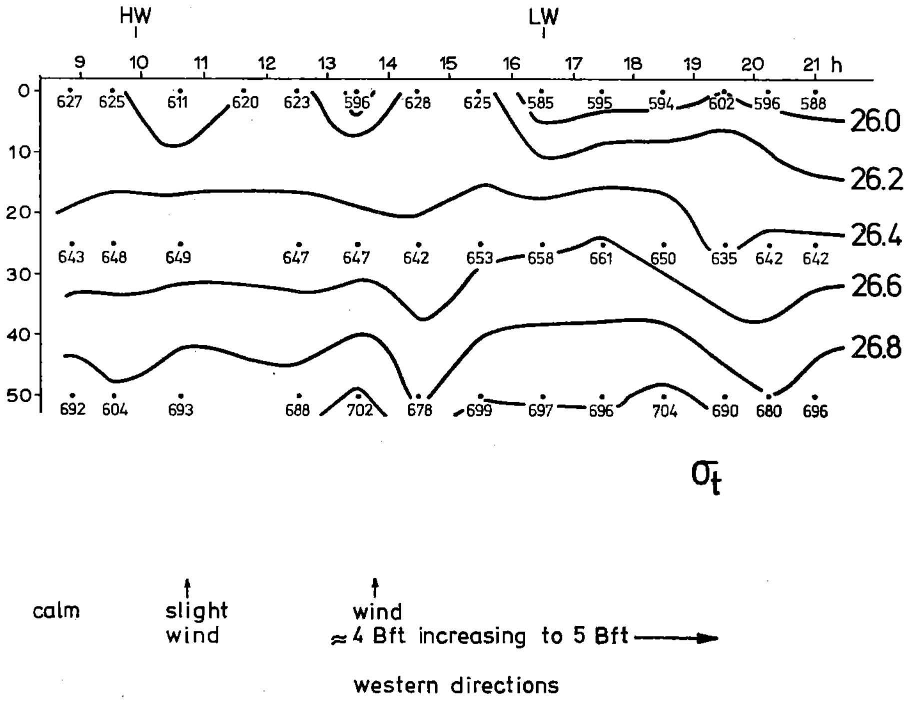


Fig. 6.27 Variation of oceanographic properties or an anchor station, 11 July 1962.



surface temperature after low tide at 11 h, on the other hand, is probably an effect of the diurnal heat exchange.

The oxygen values at the surface do not show specific variations that might be ascribed to either advective processes or a diurnal change in the relation between oxygen production and consumption by biological processes. The deeper layers do not clearly exhibit a tidal variation; a diurnal variation might be present.

Near the bottom no water of $\sigma_t = 27.0$ was observed, although at this position, near the entrance of the ria, oceanic influences might be expected; apparently there was no important inflow of North Atlantic Central Water at the time of observation. At the surface low density water was observed at the times of low water, but also, in the beginning of the series of observations, during flood. As there are no data on the wind from this anchor station, we cannot be sure about the surface circulation.

6.14 Distribution of water properties in some cross-sections

An idea of the changes occurring in the distribution of the water properties during the course of a tide is also given by two series of observations made at successive

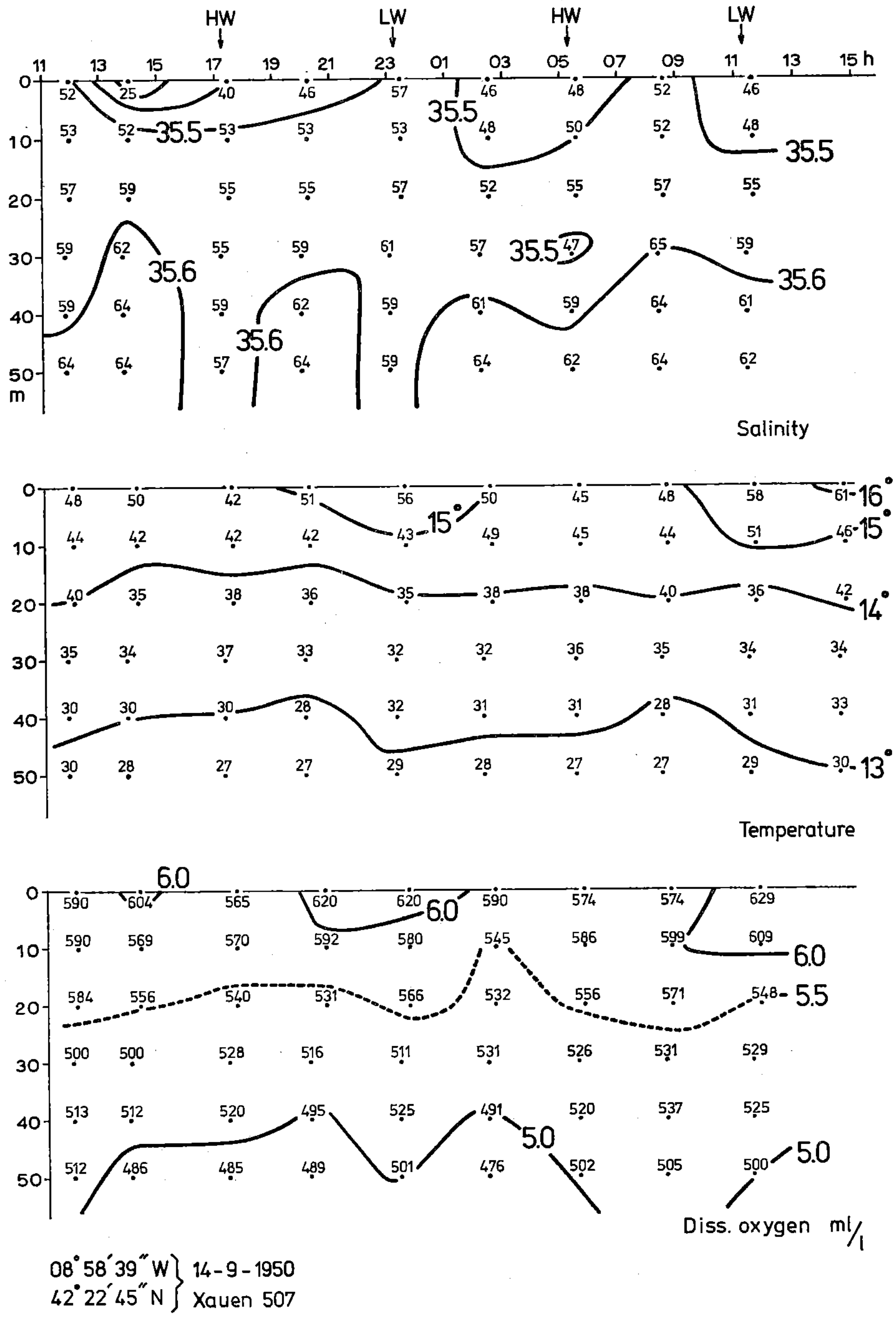
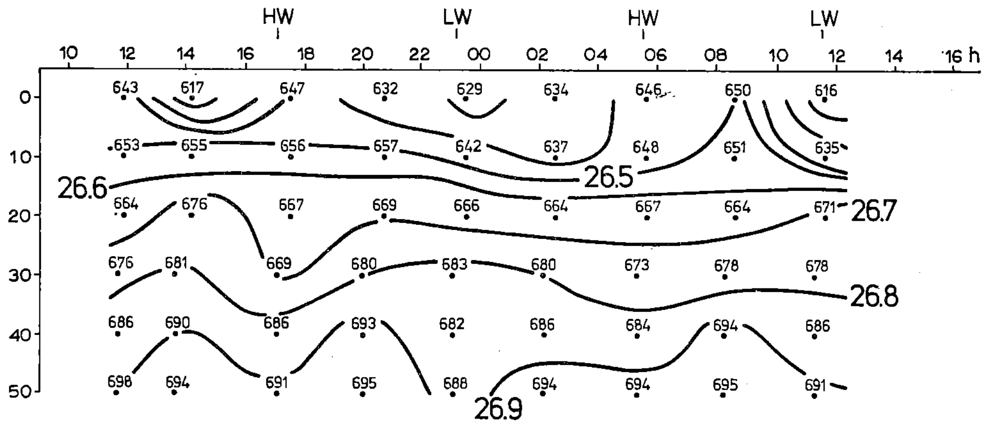


Fig. 6.28 Variation of oceanographic properties 14/15 Sept. 1950 (data 'Xauen').



intervals during the course of a day, on July 21 and July 23, 1964 (see figs. 6.29 and 6.30 respectively). In both cases the first group of observations was made some hours before high water, the second group around the time of high water and the third group during falling tide.

The observations of course could not be made simultaneously, they were worked off in a sequence that is from left to right in the figures (from west to east). This may lead to a certain deviation of the profiles shown from a true synoptic situation, especially because the period of internal oscillations is estimated to be about 1-3 hour. Nevertheless the results are interesting, as they show rather conspicuous variations. There are changes in the slope of the isopycnals as well as in the area occupied by water of a certain salinity, temperature or density.

It may be seen that for the observations of July 21 during the flood a greater part of the deeper layers is occupied by water of $\sigma_t = 27.0$, but that during the time of high water this water body is considerably reduced and that it is completely absent during the ebb. This can be taken as an indication that the inflow of oceanic water is not a continuous process, but that it may be subject to interruptions.

The observations of July 21 and 23 both show changes in the inclination of the isopycnals which might indicate changes in the deeper currents but as the time difference between the successive observations makes a good estimate of the relative slope impossible and furthermore accelerations could cause important deviations from the geostrophic situation a qualitative discussion of these current variations appears impossible.

However, in the following section we will attempt to deduce the mean depth variation of the net current from the mean density distribution.

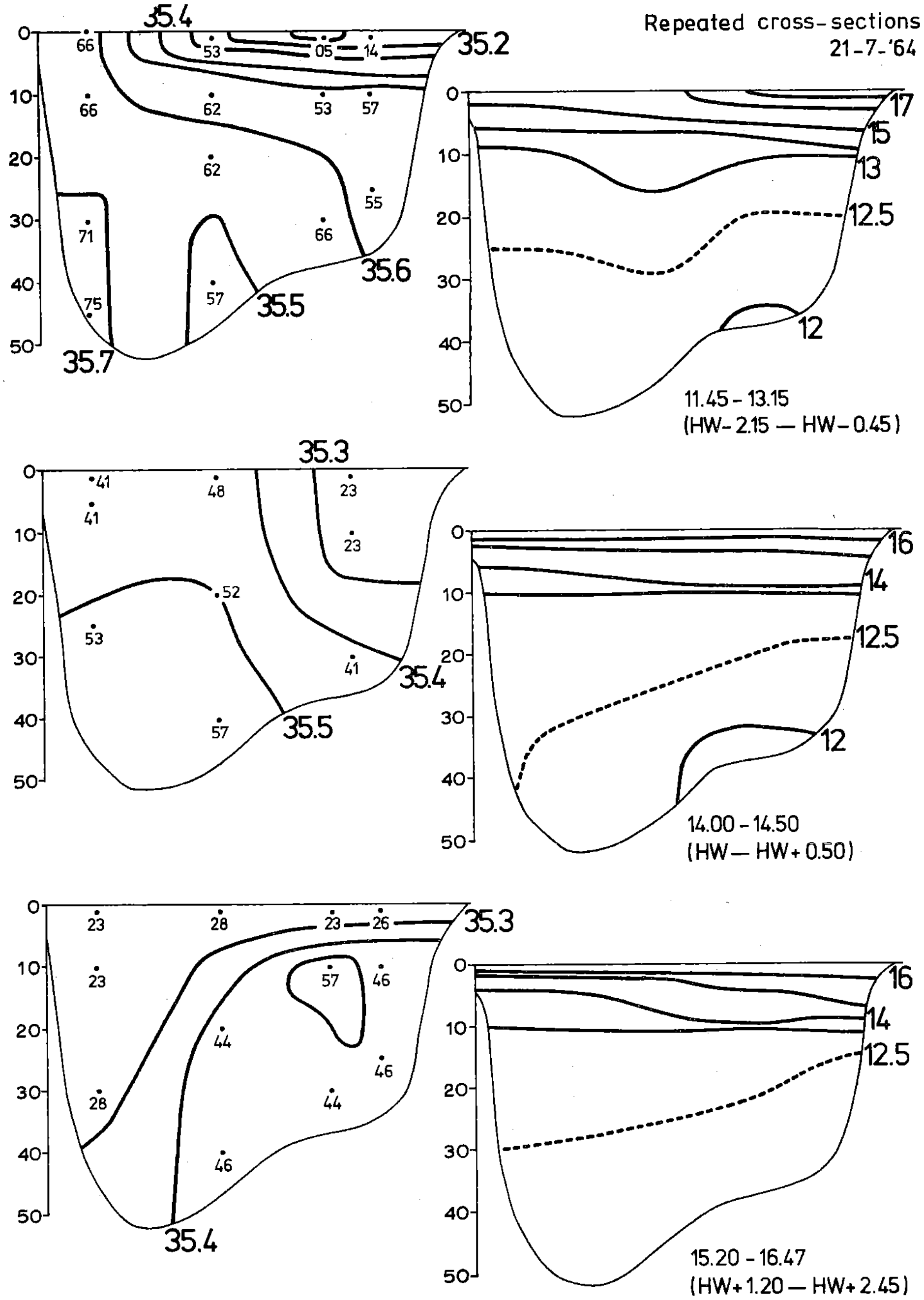
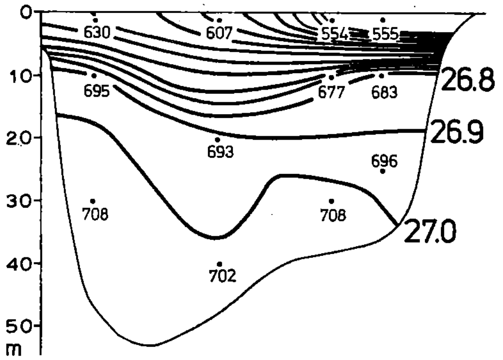
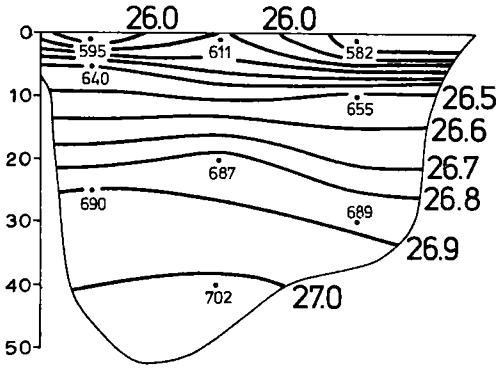
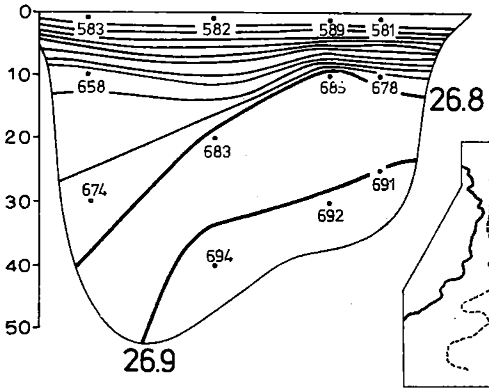


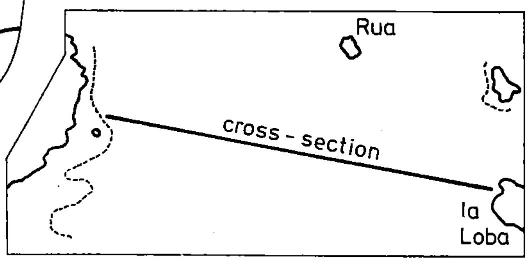
Fig. 6.29 Variation of oceanographic properties on cross-section, 21 July 1964.



Wind N
increasing
6 $^{m}/_{sec}$ → 7 ½ $^{m}/_{sec}$
(4 Bft)







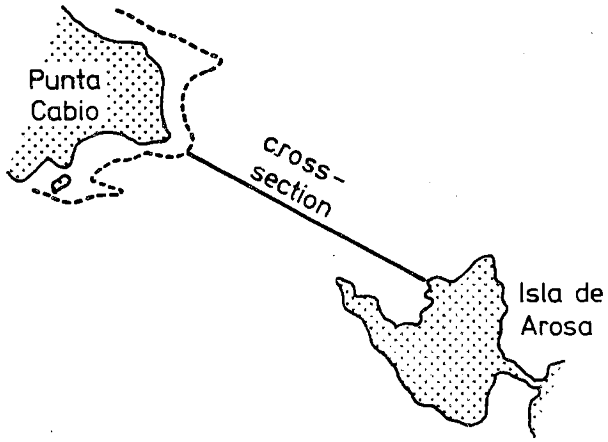


Fig. 6.30 Variation of oceanographic properties on cross-section, 23 July 1964.

6.15 Geostrophic approximation

The assessment of currents by means of the geostrophic approximation is a normal practice in circulation studies of the oceans. For a coastal area, however, this method usually is not recommended, because the basic assumptions, viz. negligible effects of friction and acceleration terms in the equation of motion, are invalid. However, in a study of the currents in the St. Lawrence estuary, FORRESTER (1970) has shown that, if observations are made in such a manner that the effect of short-period fluctuations (internal cross-sectional seiches) is averaged out, the geostrophic approximation gives the mean relative current over this interval with respect to the reference level with a reasonable accuracy.

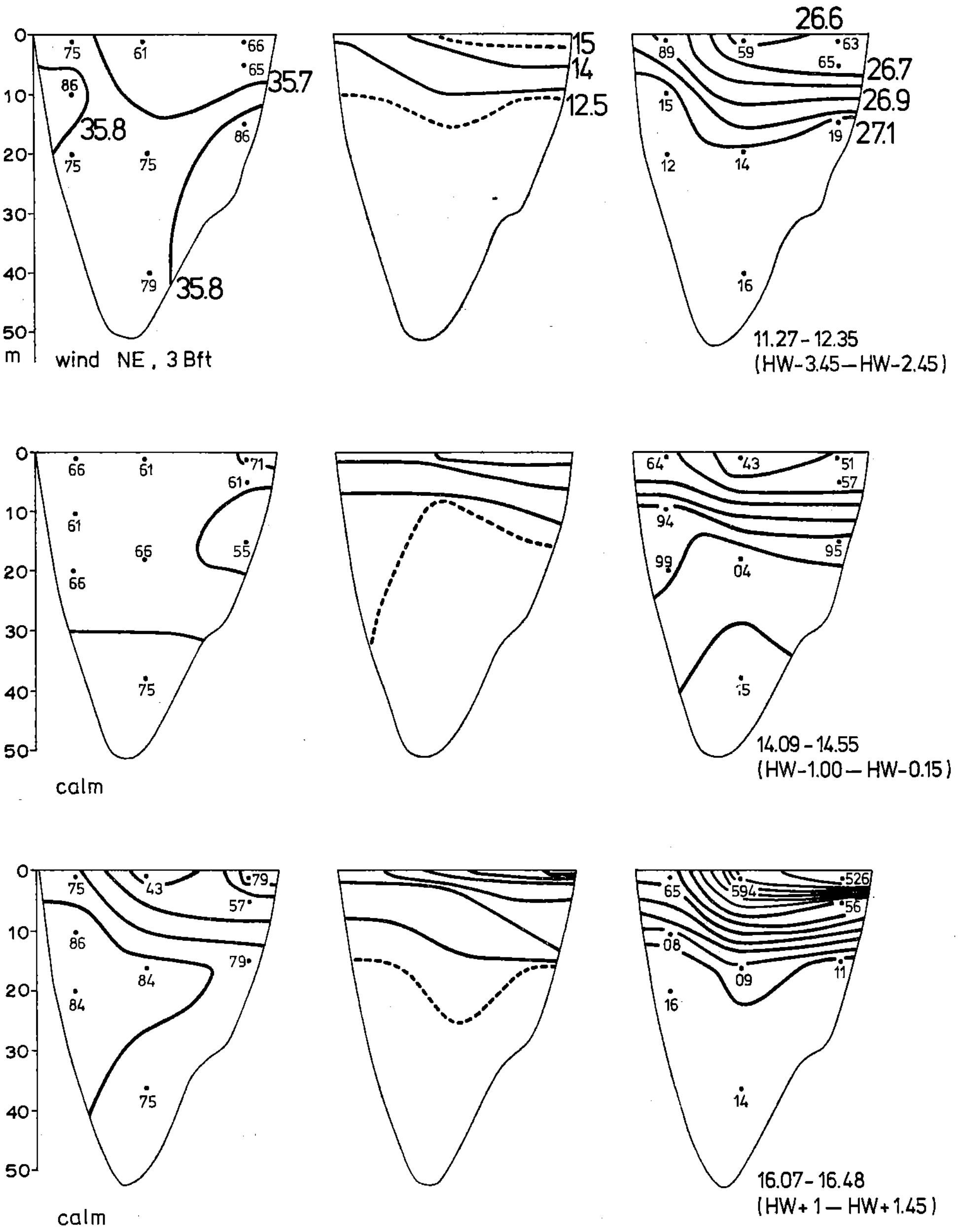
Now, the Ría de Arosa is much smaller and shallower than the St. Lawrence estuary, and friction and field acceleration will be more important. Nevertheless, a geostrophic calculation based upon the mean pressure distribution of a relatively uncomplicated part of the main channel, such as at the cross-section worked-up on July 21, 1964 (fig. 6.29) may give at least some additional information on the current pattern.

The time interval in which the observations were made was symmetrical with respect to the moment of high water, which means that the contribution of the cross-sectional mean of the tidal current to the total current is fairly small. The mean pressure distribution calculated from the three series of observations gives the following pattern.

In the eastern part of the section at 0-10 m depth an inward current of 12-18 cm/s was found relative to the (unknown) current at 30 m depth, while in the western part successively an outward current of 6-17 cm/s was found at the upper 10 m and an inward current of 5 cm/s at 20 m, also relative to the current at the 30 m level.

Assuming the current at 30 m to be small (taking into account the results of the

Repeated cross-sections 23-7-'64



observations mentioned in section 6.11), the values given above indicate the order of magnitude of the in- and outflowing residual currents.

A geostrophic calculation for the transverse section of July 23, 1964 (fig. 6.30) gives no acceptable results. The topographic situation is more complicated and this method will fail sooner because of the influence of the acceleration terms.

There is a suggestion of an inflowing surface current in the eastern part of the outer ria, as also was inferred from the salinity distribution, and of a reversal of the current direction in the western part somewhere between 10 and 20 m depth.

6.16 Oceanography of some special regions

The foregoing paragraphs largely dealt with the main channel of the Ría de Arosa. Because of a necessary schematization interesting features of some special regions had to be neglected. In the following paragraphs certain aspects of several marginal regions will be discussed, as in some way or other they show special points of interest.

This discussion has to be more or less superficial, as in the programme of observations the emphasis has been laid upon the main channel. As even there the observations not always reached out to reveal all particulars of the circulation pattern, the oceanographic situation in other parts of the ria that have been sampled less frequently only may be discussed in a rather general way. In several places the situation deserves a more intensive exploration than could be accomplished here.

The following regions will be discussed:

a. Ensenada de la Puebla de Caramiñal (Bay of Puebla de C.)
b. Channel of El Vado, east of Isla de Arosa
c. Basin in front of the Rio Umia ('Grove deep')
d. Ensenada del Grove (Bay of El Grove).

6.17 The Ensenada de la Puebla de Caramiñal

A number of oceanographic observations at different depths was made in this area on July 27, 1964.

The surface salinity as observed on that occasion is shown in fig. 6.31. The most important point is the increase of salinity from the mouth of this bay towards its head. Apparently the run-off from the small rivulets that discharge into this bay is of no importance for the salinity distribution.

The variation with depth of the water properties is given in the form of a longitudinal section (fig. 6.32). The salinity near the bottom slightly decreases towards the inner parts of the bay and the vertical differences of the salinity are considerably smaller in the inner parts than at the mouth. The temperature at the surface increases towards the inner parts and this shows that upwelling is not responsible for the

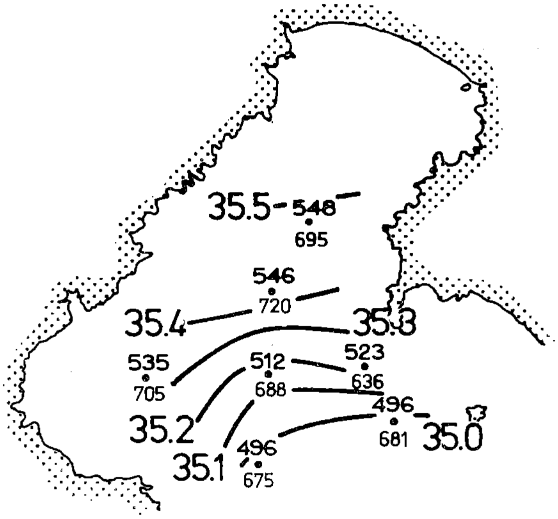


Fig. 6.31 Surface salinity, Ensenada de la Puebla de Caramiñal (27 July 1964).

higher salinities that are found at the surface near the head of this embayment. The density distribution anyhow is in favour of an inward current at the surface and indeed strong inward currents have been observed on 30 July 1963 while a southwesterly wind was blowing, during the ebb period (fig. 6.23).

The density distribution given in fig. 6.32 also shows that high-density water with the density of oceanic water is found at the entrance of the bay and the gradual decrease of the density of the bottom water towards the inner parts suggests an inward movement along the bottom. The vertical differences in the density gradually diminish towards the inner parts, probably by vertical mixing. The effect of evaporation may also play a part, although it is thought not to be the most important one for the observed increase of salinity in the upper layers would, even at an evaporation rate of 2 mm/day, lead to residence times of over 15 days, a value that appears incompatible with the observed current speeds at the entrance of the bay.

In such a situation we may assume that the surface water will sink and that the bottom water will ascend in the inner parts of the bay and that the water will flow out of the bay at intermediate depths. This kind of circulation has been described by PRITCHARD AND CARPENTER (1960) for Baltimore Harbor, a marginal embayment of the Chesapeake Bay. The Ensenada de Caramiñal, however, is much shorter ($2\frac{1}{2}$ miles against about 9 miles for Baltimore Harbor) and therefore this type of circulation is less clearly developed. Furthermore it is possible that for stronger northerly winds upwelling may occur, although for the case studied here wind conditions were also favourable for upwelling (6-7 m/s from northern directions).

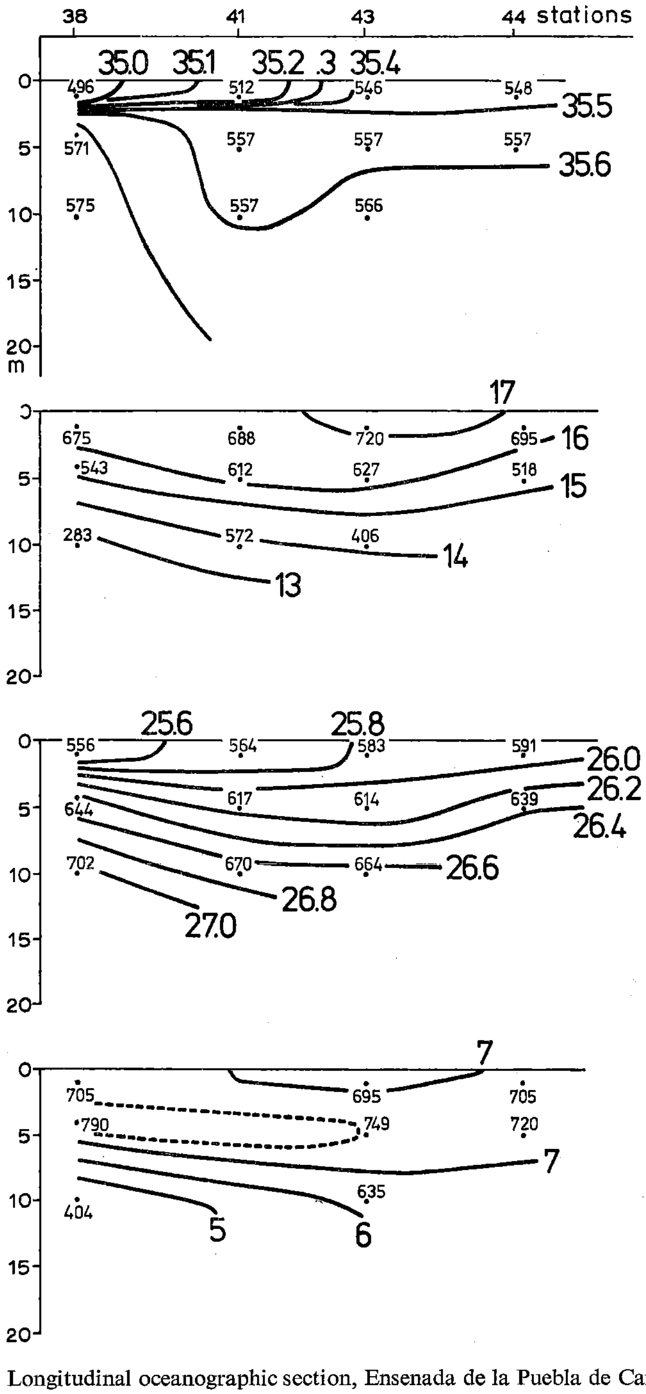


Fig. 6.32 Longitudinal oceanographic section, Ensenada de la Puebla de Caramiñal (27 July 1964)

6.18 The channel of El Vado

The area east of the Isla de Arosa is of special interest because here a connection exists between the southeastern part of the ria, which is more or less influenced by the water from the Umia river, and the inner part of the main channel, which is under influence of the outflow of the Ulla and of the upwelling process described in previous sections.

The first impression is that because of the small depths in the area east of Isla de Arosa and of the barrier formed by the sand spit of El Vado the rate of exchange will be rather small. The current data given in fig. 6.33 show that the currents in the channel near El Vado are often very weak. We shall discuss in detail the current measurements made at this place. These observations have been combined in a composite figure which shows the variation of the current with the two predominating wind directions. On July 10 and 29, 1963, observations were made with southwesterly wind, which observations have been combined in the upper curve of fig. 6.33. Most of the time the current was less than 5 cm/s, only around the time of low water current

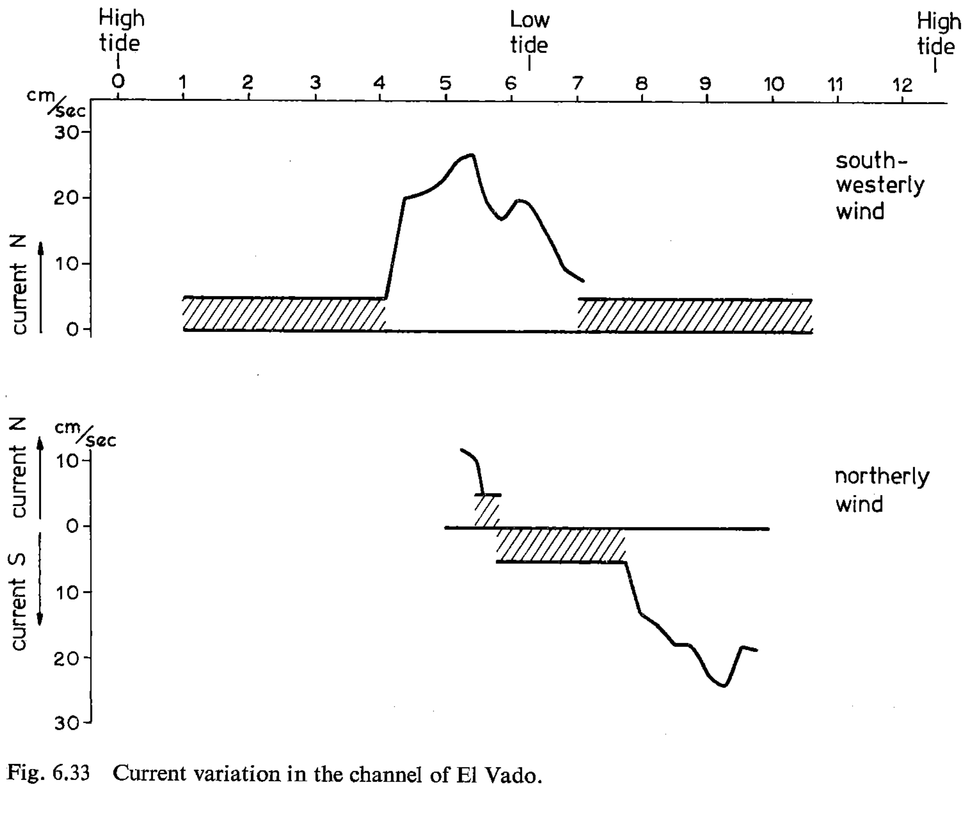


Fig. 6.33 Current variation in the channel of El Vado.

velocities of greater importance were observed, apparently because of the decrease of the aperture of the channel, as a large part of the sand spit then is above sea level.

The velocities observed with northerly wind, on July 28, 1964, show a relatively important southerly current during the flood. This probably reflects the influence of the wind. However, during the end of the ebb period, in spite of the wind direction, weak northerly currents were measured. Apparently, around the time of low water there is a tendency for a northerly current that is sufficiently strong to overcome the opposed wind stress. One might assume that apparently there is at that time a difference of water level between the basin in front of the Umia and the main inner part of the channel of the ria. The northerly set of the ebb current east of Isla de Arosa observed by GÓMEZ GALLEGO (1971) supports this observation.

Summarizing these data, we may state that during the time of low water a northerly current exists, the strength of which depends largely on the prevailing wind. But as the flow is constricted by the small aperture of the channel at that time, the total water exchange is of relatively small importance. During a northerly wind, however, there is during the flood period a southward movement of water that may be of some importance, as at that time the water level is higher and the depth and width of the channel are greater.

6.19 The 'Grove deep'

The bay in front of the estuary of the Umia river forms a more or less separate basin of the ria. The surface water is influenced by the outflow of this river. Also the exchange with the shallow bay of El Grove (see section 6.20) influences the conditions of the surface layers. The centre of the bay is about 20 m deep, this deep part ('Grove deep') is separated from the greater depths in the main channel of the ria by a sill between the Peninsula of El Grove and the Isla de Arosa. This sill has a depth of less than 10 m. The properties of the water beneath sill level were found to be different on both sides of the sill. An example is given in fig. 6.34, which shows the vertical distribution of the temperature obtained from a series of bathythermograph measurements made at a section from west to east over the sill on August 4, 1962. The water of the deeper layers within the basin was about 1 °C warmer than the water outside at the same depth.

In such a case of a relatively deep basin separated from the main body of the water by a shallow sill the renewal of the water beneath sill level may be hampered. Besides the conditions found here, where warming-up of the water and run-off from rivers give a stable stratification of the water mass, are in favour of stagnation of the deeper layers. Renewal of water in such a barred basin can occur by vertical mixing or by a process by which water flows over the sill into the basin. The latter implies that at the oceanic side of the sill water is present with a higher density than in the basin inside. Such a rising of water with a high density upto sill depth may occur during periods

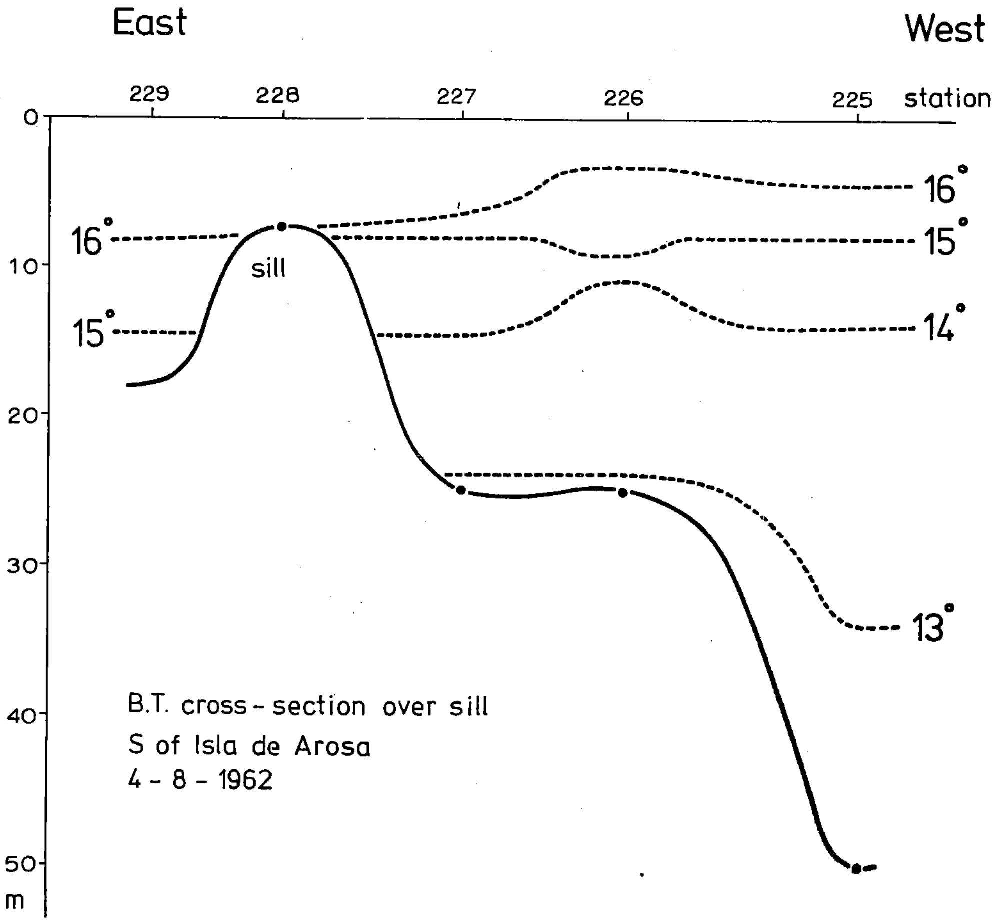


Fig. 6.34 Temperature section Grove deep (4 August 1962).

of intense upwelling, and may happen only occasionally. This process may occur several times, but at last the water within the basin would attain such a high density that renewal would become very rare indeed, if no vertical exchange with surface water would gradually decrease the density in the basin again. In a relatively shallow basin such as we find here this vertical mixing will be more effective than in cases with a much deeper basin. The frequency of the renewal of the deep water is of importance for the conditions in the water, as with a low frequency there may be an appreciable decrease of oxygen concentration, due to oxygen consumption.

Several times observations were made in the basin. The results of the observations are given in the form of vertical profiles of salinity, temperature, density and dissolved

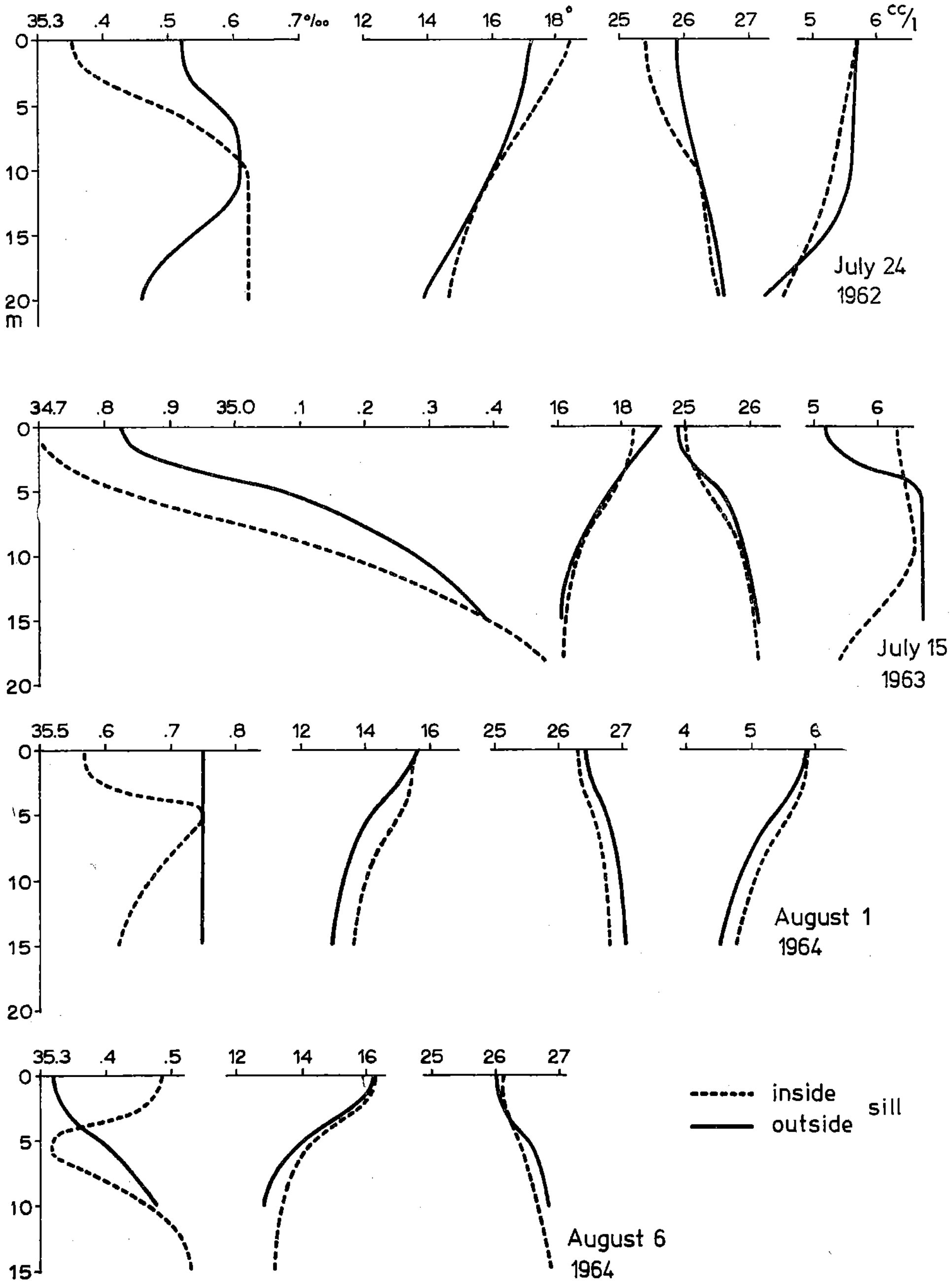


Fig. 6.35 Vertical profiles observed on various occasions in the Grove deep.

oxygen (fig. 6.35). In the same figure similar data of stations just outside the sill are presented. The conditions that have been found in the basin are also shown in table IX.

TABLE IX. *Oceanographic data from the 'Grove Deep'*

	S	t	σ_t	O_2
July 24, 1962	35.62 ‰	14.67 °C	26.54	4.49 ml/l
August 4, 1962		ca. 15		
July 15, 1963	35.48	16.12	26.10	5.35
August 1, 1964	35.62	13.68	26.75	4.76
August 6, 1964	35.53	13.22	26.78	no obs.

There is a tendency towards higher densities later in summer, but differences between the different years may of course be of influence.

The observations of July 24, 1962, and July 15, 1963, show situations in which the water in the basin had higher σ_t values than the water outside the basin at sill depth. This means that at that moment no overflow of the sill occured. On August 1, 1964, the density outside the basin at sill depth was higher than in the basin, and the situation therefore was favourable for overflow. If this process would have happened for a sufficiently long time interval the water in the basin would have reached the same temperature, salinity and oxygen content as the water outside at sill depth, that is:

$$S = 35.75\,‰, \quad t = 13.3\,°\text{C}, \quad \sigma_t = 26.9, \quad O_2 = 4.7\,\text{ml/l}$$

However, the values found some days later, on August 6 were different, as is seen from the table given above. The conclusion is either that the overflow was not sufficient for a complete renewal, or that the properties of the deep water had changed in the meantime by vertical mixing.

From the density distribution found on August 6 it follows that some inflow might occur at that time, but not down to the bottom of the basin.

These observations lead to the conclusion that an intermittent overflow into the basin occurs. The oxygen data do not suggest that the frequency of this overflow is low. It may be, however, that in certain years or perhaps later in summer the overflow is less frequent and that low oxygen values occur.

6.20 The Ensenada del Grove

In several respects the Ensenada del Grove takes a special position in the oceanographic situation of the ria. This bay for the major part consists of tidal flats and is connected with the deeper parts of the ria by a relatively narrow channel between the islands Toja Grande and Toja Pequeña. No appreciable freshwater inflow takes place in this bay.

From the depth distribution given in the nautical chart of that area edited by the Spanish Hydrographic Service it may be estimated that the total volume of water within the bay varies between about 26×10^6 m^3 at high tide and about 5×10^6 m^3 at low tide. So about 80 % of the volume water beneath the high water level flows into and out of the bay during a tidal period.

The distribution of temperature and salinity as observed during the ebb on July 31, 1964, is shown in fig. 6.36 for a section from the mouth towards the head of the bay.

There is a gradual increase of temperature from the mouth of the bay inwards. The difference between surface and bottom temperature is about 0.4 °C. The salinity at the bottom shows a tendency to decrease towards the head of the bay, but at the sur-

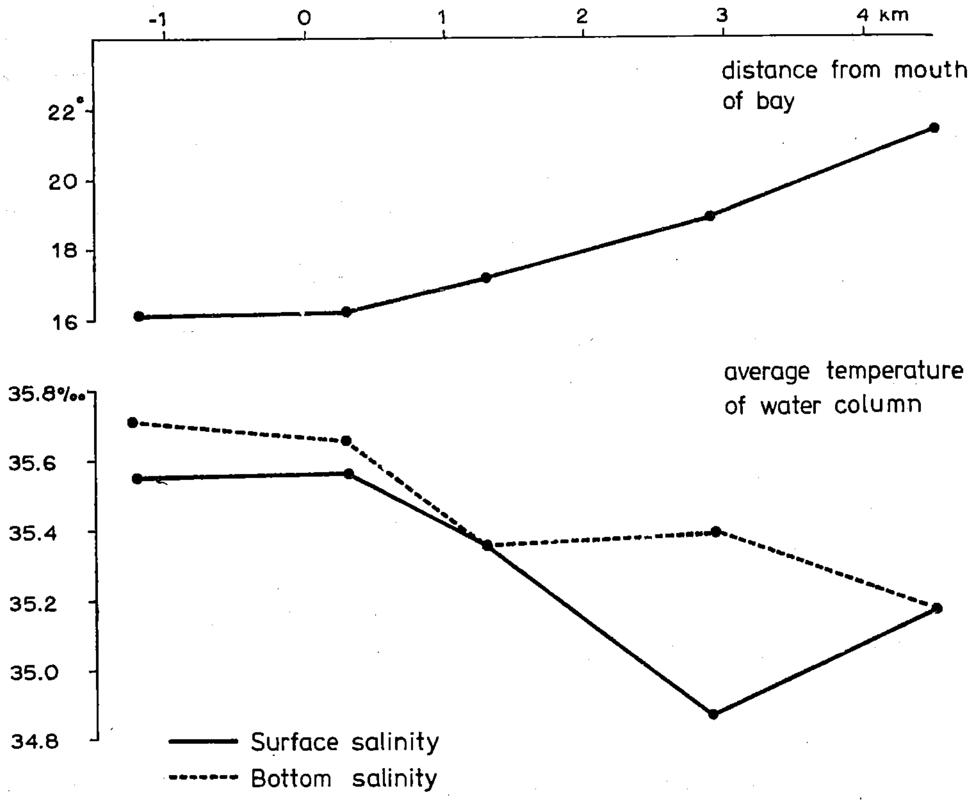


Fig. 6.36 Longitudinal variation of salinity and temperature in the Ensenada del Grove, 31 July 1964.

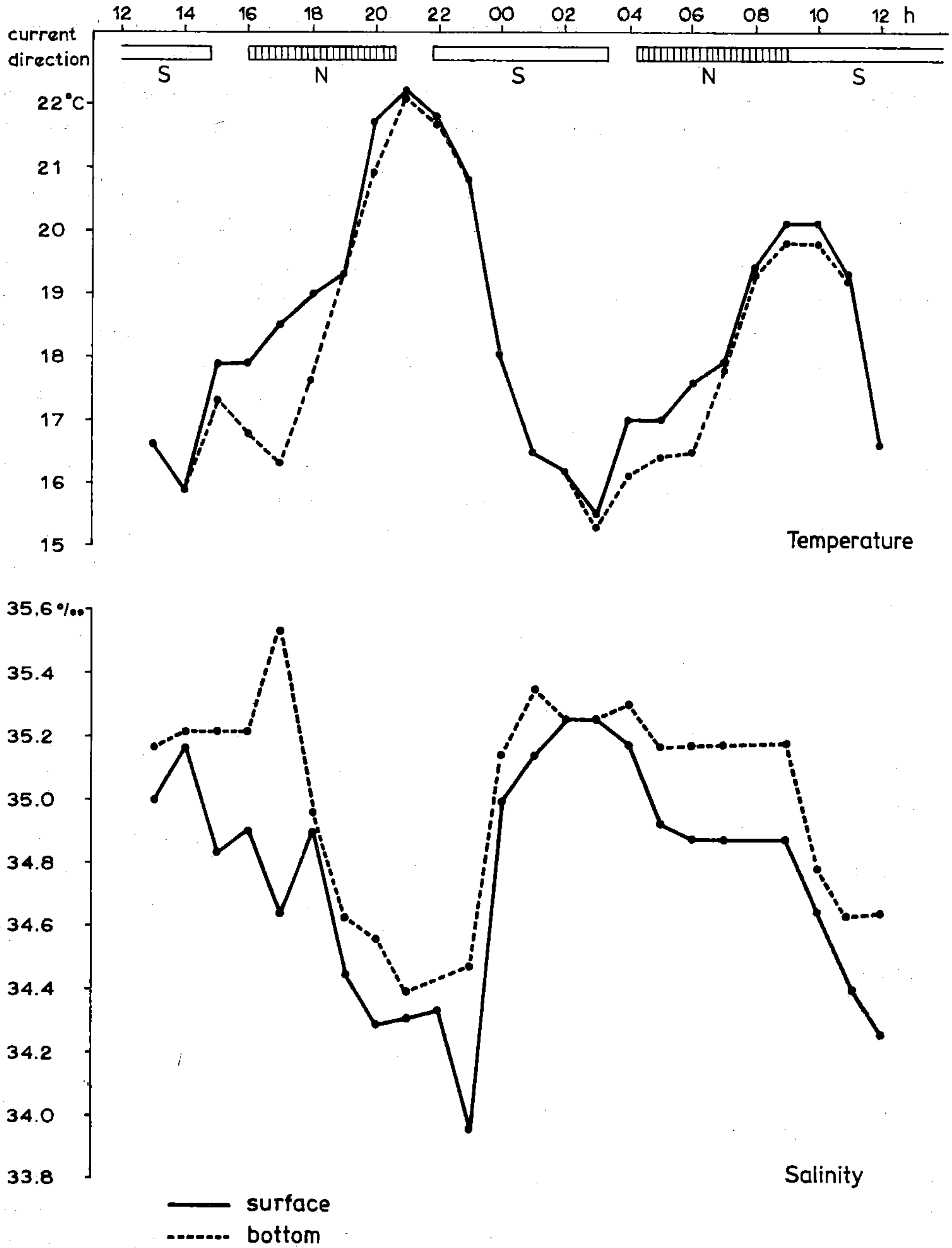


Fig. 6.37 Variation of oceanographic properties on an anchor station near La Toja, 19 and 20 Juli 1963.

face there is a distinct minimum at some distance from the head. Here we probably find water that during the end of the previous ebb period came out of the Umia estuary and was pressed into the bay of El Grove during the following flood. This inflow of relatively low-salinity water was also observed passing an anchor station near La Toja during the first stage of the flood.

The high temperatures in the inner part of the bay are apparently caused by the local heating in this shallow area. The incoming radiation on a cloudless day is estimated to give a temperature increase of several degrees per day if the water depth is no more than a few metres. Of course there will also be loss of heat by exchange with the atmosphere and by back radiation, but a distinct temperature increase is quite probable, although with each tide the major part of the water is renewed.

For the study of the exchange of water between the bay of El Grove and the rest of the ria the observations made on an anchor station in the connecting channel near La Toja are of interest. These observations were carried out on 19 and 20 July 1963 over a period of 24 hours. The results are given in fig. 6.37.

The temperature variation runs parallel to the variation of the current: during the flood (southerly current) temperature decreases, during the ebb (northerly current) temperature increases.

The salinity variation is not completely similar to the temperature variation. There was a continued decrease of the surface salinity during the beginning of the flood on both occasions; near the bottom only during the beginning of the second flood period a continued decrease is noted. As has been said before, this water probably came from the Umia estuary and reached the area in front of the channel of La Toja during the previous ebb period.

Because during the ebb current no such low salinities are observed, we may conclude that the waters inside the bay of El Grove become fairly well mixed during a tidal period. This mixing with a part of the outflow of the Umia river may be responsible for the absence of a marked tongue of low-salinity water in the adjoining sea area (see fig. 6.1).

6.21 Additional observations from de Ría de Arosa

Apart from the data of GÓMEZ GALLEGO (1971) there are some further data on the properties of the water in the Ría de Arosa from cruises by the Spanish research vessel 'Xauen'. See also data from the anchor station made on September 14-15, 1950, in section 6.13. Besides, there are two series of observations in the harbour of Villagarcia, on 11-12 September 1949 (INSTITUTO ESPAÑOL DE OCEANOGRAFÍA, 1955) and 25-26 August 1952 (INST. ESPAÑOL DE OCEANOGRAFÍA 1961), and a series of stations made on September 22 and 23, 1949. The data of Villagarcia do not give new aspects. However, the observations of September 1949 made throughout the ria show a vertically rather homogeneous water mass from the surface down to the bottom, with temperatures not falling below 17.8 °C. There are some incongruities in the data (instabilities, apparent printing errors) so we cannot be completely sure of the values given. But, as the data of the anchor station of September 1950 show a definite stratification while on the other hand the data given by ANADÓN ET AL. (1961) show that in the Ría de Vigo similar cases of homogeneity have been found in October, we may just note that these data (if reliable) suggest that in autumn the stratification may sometimes be destroyed.

7. Oceanographic situation in the winter of 1964

7.1 Distribution of temperature and salinity

During the short winter campaign from January 28 till February 7, 1964, only a limited number of observations could be made. Accidental influences on the oceanographic situation may therefore obscure more significant effects. Charts showing

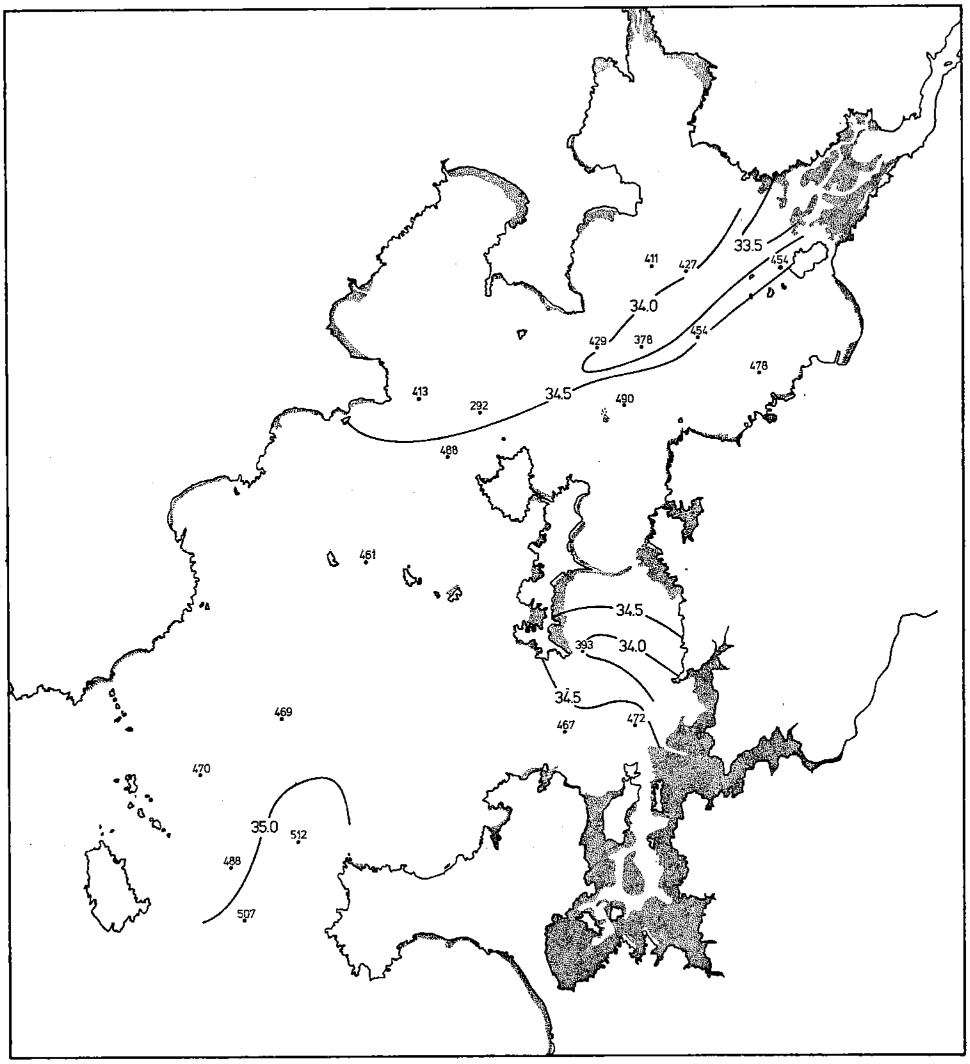


Fig. 7.1 Salinity distribution at the surface, Ría de Arosa. Winter 1964.

salinity and temperature of the surface waters are presented in fig. 7.1 and 7.2. These charts were drawn using the individual observations, and not the mean values of a larger number, as is the case with the charts presented in fig. 6.1 and 6.2. As some of the observations gave more or less extreme values that did not appear to fit in the whole pattern, these observations were given less weight. Therefore a certain degree of subjectivity has to be taken into account when considering the charts.

The surface salinity shows a certain decrease with respect to the summer situation.

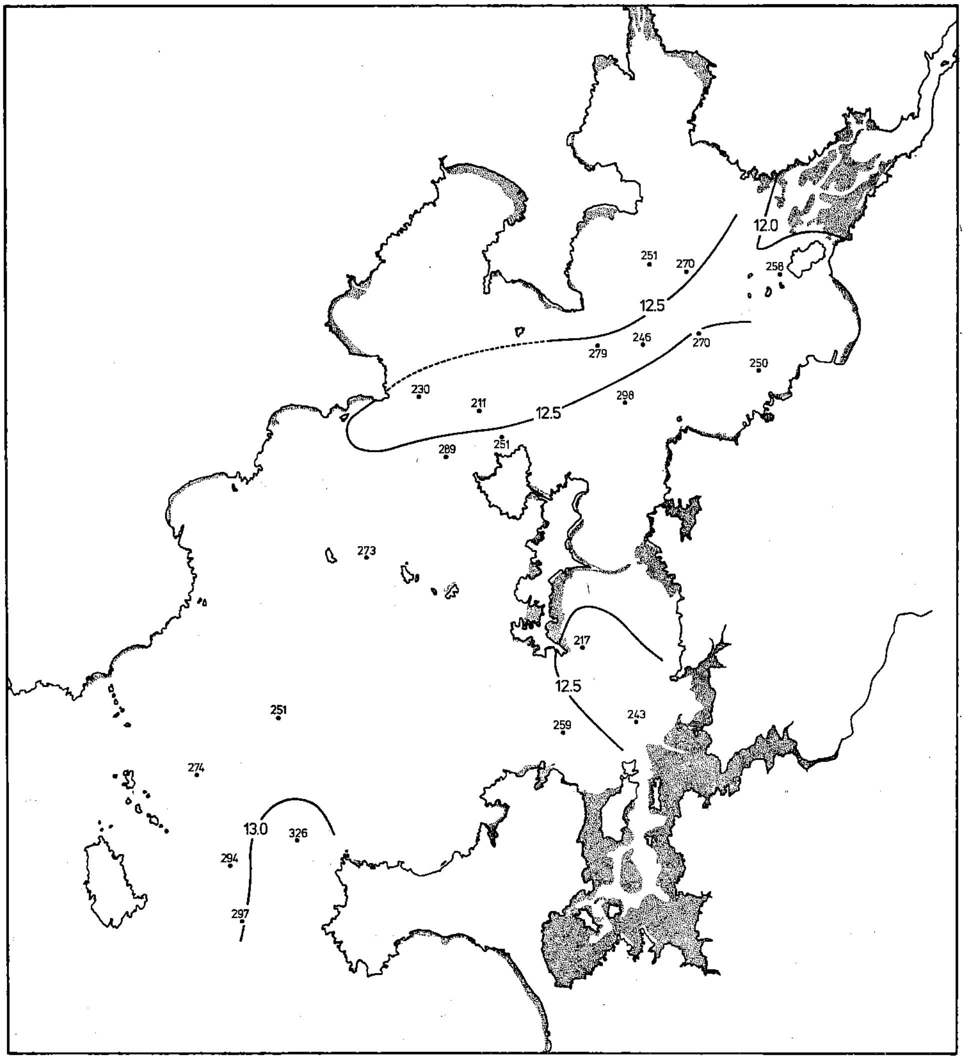


Fig. 7.2 Temperature distribution at the surface Ría de Arosa.

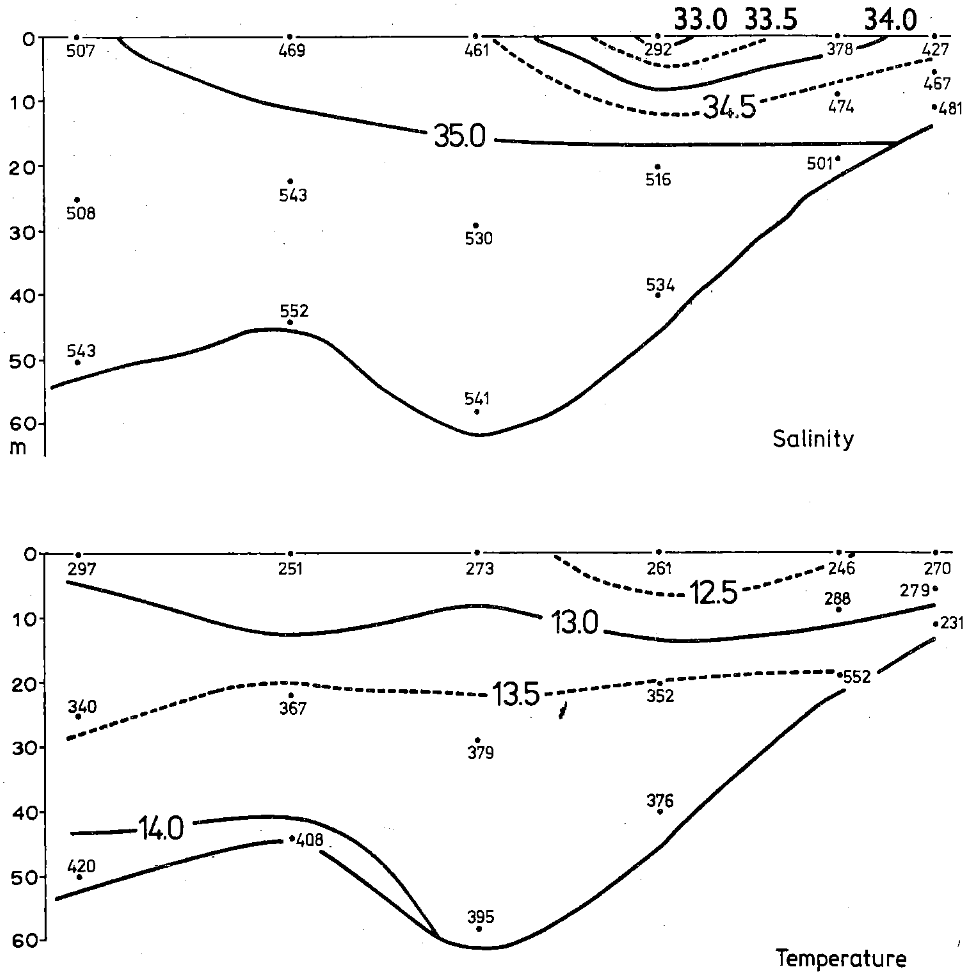
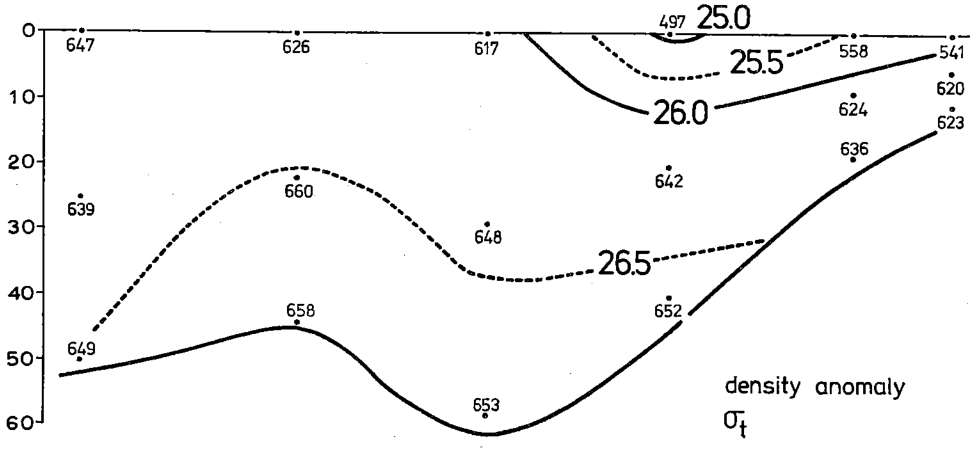


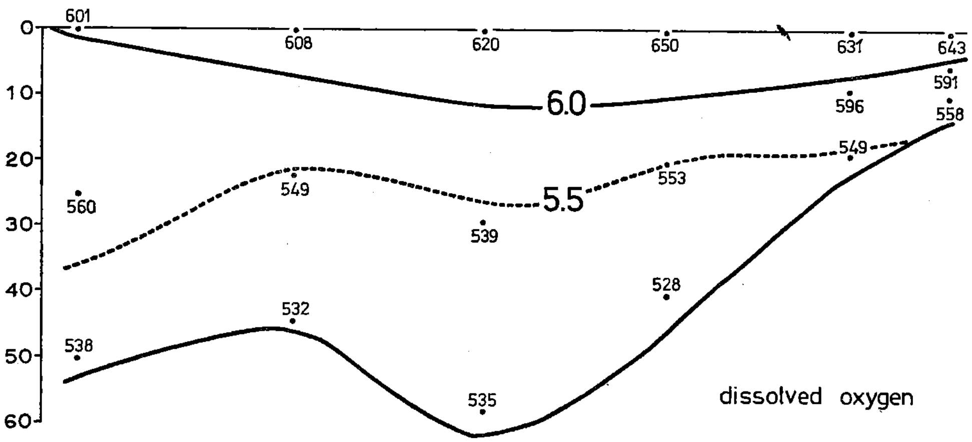
Fig. 7.3 Longitudinal oceanographic section, Ría de Arosa. 30 January 1964.

The difference is of the order of 0.5 $^0/_{00}$. The salinity pattern, however, has many points in common with the summer situation.

The surface temperature chart shows rather small differences between different places of the ria. The major part of the ria has temperatures ranging between 12.5 °C and 13 °C. This is quite different from the summer situation, where the temperature ranged from 15 °C till 18 °C.

A further illustration of the oceanographic situation is given in the form of a number of sections, showing the distribution of temperature, salinity density and dissolved oxygen with depth.





A longitudinal section of 6 stations made on 30 January is shown in fig. 7.3. There are some irregularities, especially a minimum of salinity and temperature at the surface that, contrary to expectation, is not observed in the inner part of the ria, but at some distance from the mouth of the river Ulla. This may be caused by non-synchronous sampling, but it may also be the effect of irregularities in the outflow. Apart from these irregularities, some points of interest may be noted. First, the occurrence of low temperatures at the surface, decreasing with decreasing salinity, and of higher temperatures near the bottom. Furthermore the rather small variations of the density of the deep water are noteworthy. The oxygen content at the surface is higher than

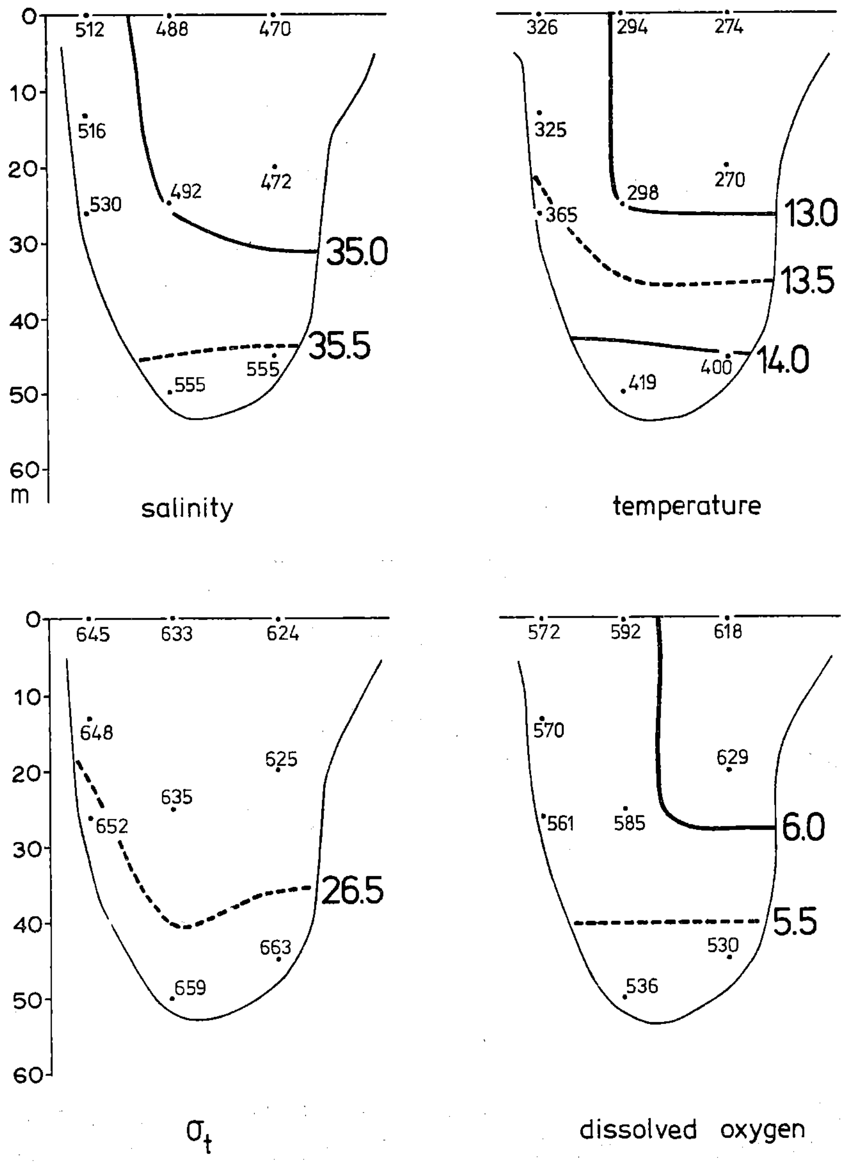


Fig. 7.4 Transverse section, Ría de Arosa. 31 January 1964.

at the bottom, as should be expected; there is an oversaturation at the surface and the bottom waters are slightly undersaturated. All transverse sections show the same general features, with relatively low temperatures and salinities at the surface and especially on the western side of the ria. A transverse section of the mouth of the ria, is given in fig. 7.4.

A further comparison with the summer situation is possible by means of the vertical profiles of salinity, temperature, σ_t and dissolved oxygen for stations that are at the same positions as the positions A, B and C, for which the average vertical profiles are given in fig. 6.4 of section 6.1.

The corresponding profiles for the winter of 1964 are shown in fig. 7.5.

The vertical temperature gradient is less steep than in summer. The bottom temperatures are higher than in summer. This reflects the absence, or at least the weakening, of the process of upwelling, by which cold water is lifted up to over the continental edge.

The vertical differences of the salinity are more important during winter than during summer. Although during the winter campaign the weather was rather dry, and the river run-off was estimated to be only twice that of the summer, the surface salinities were below those of the summer. This decrease of salinity is, however, not sufficient to compensate for the increase of density caused by the cooling of the water. Therefore the density differences are smaller than in summer, which means a weakening of the vertical stability of the water column. Finally, the oxygen values show smaller vertical differences, because of a small decrease of the oxygen content of the surface water and a larger increase of the values in the deeper water in comparison with the summer situation.

Still, notwithstanding these differences, the winter profiles have a point in common with the summer situation, viz. the increase of the vertical stability from the mouth of the ria inwards and, owing to the smaller depths at the head, the stronger vertical gradients as compared with the more oceanic parts of the ria.

The latter point indicates that under the conditions of our winter campaign the circulation may have been not much dissimilar to the summer circulation; in other words: it may also be described as a two-layered pattern of flow. But whether upwelling can have taken place too is not certain, as the small vertical differences of the water properties make it difficult to detect such a process.

7.2 Characteristics of the water masses in winter

The observed data were plotted in a temperature-salinity diagram in order to study the water masses that were present in this period and to investigate the changes that occur in these water masses by different processes.

In fig. 7.6 the data from observations below a depth of 20 m are presented. Here the water is not so much influenced by the river run-off and the exchange of heat with the

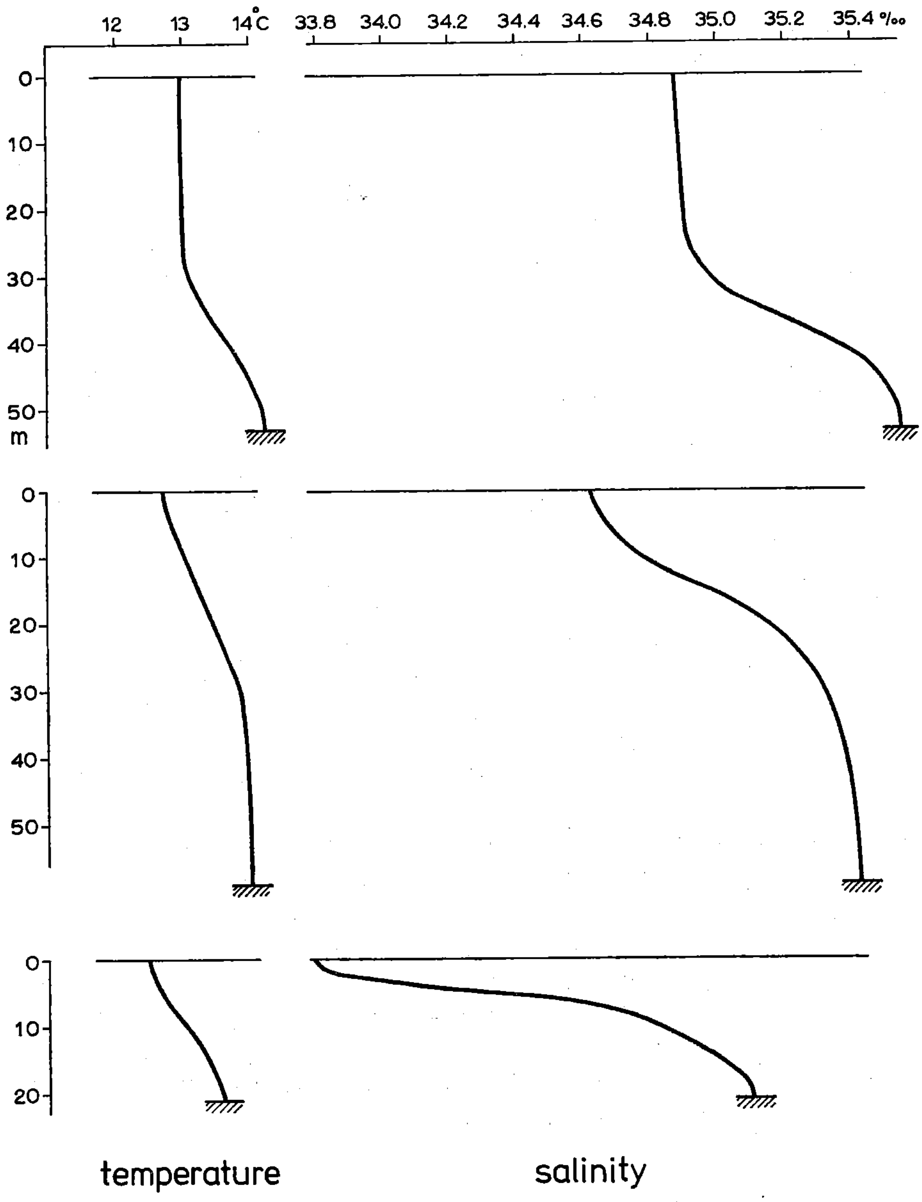
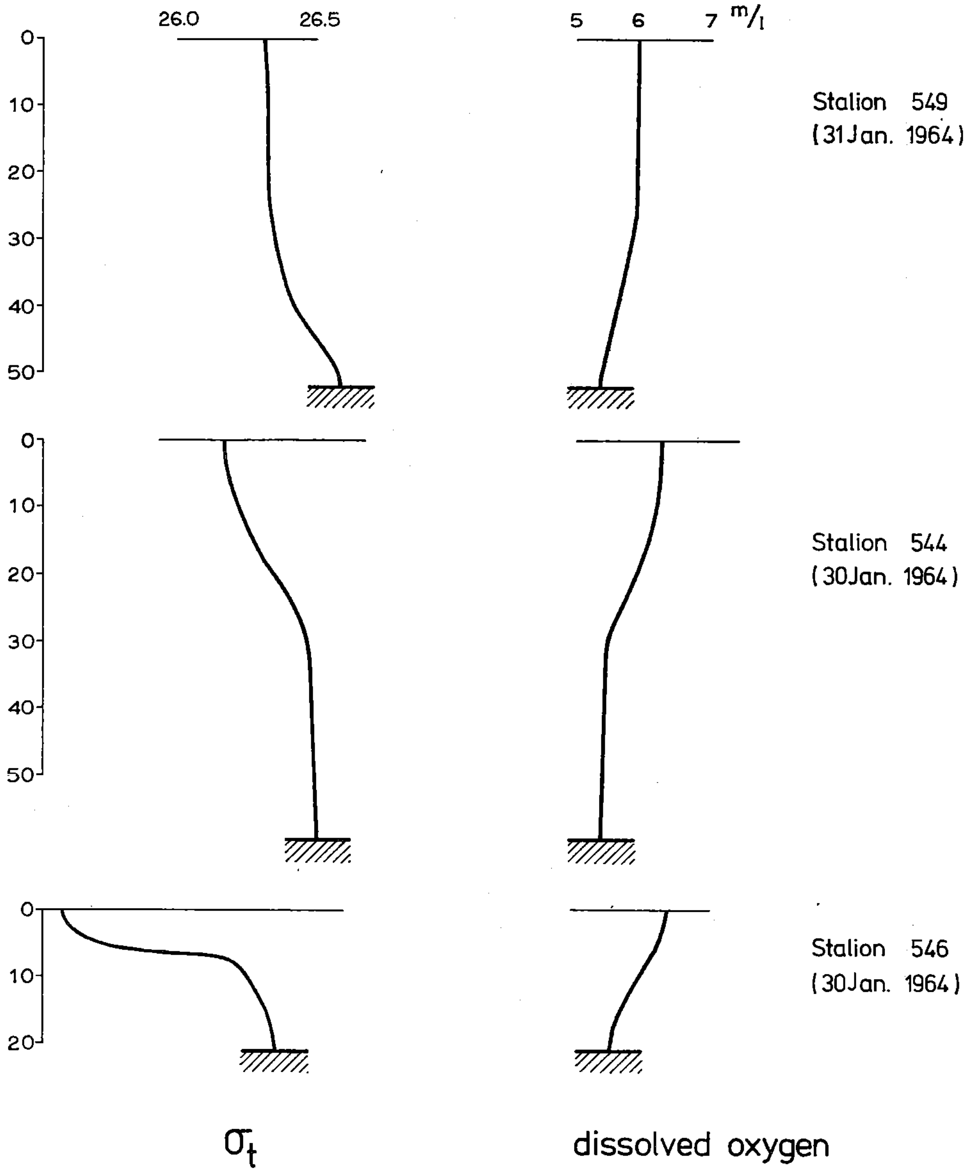


Fig. 7.5 Vertical distribution of temperature, salinity, density and oxygen content at three points in the Ría de Arosa. Winter 1964.



26.0
26.5
5
6
7 m/l
0
10
20
30
40
50
Stalion 549
(31 Jan. 1964)
Stalion 544
(30 Jan. 1964)
Stalion 546
(30 Jan. 1964)
σt
dissolved oxygen

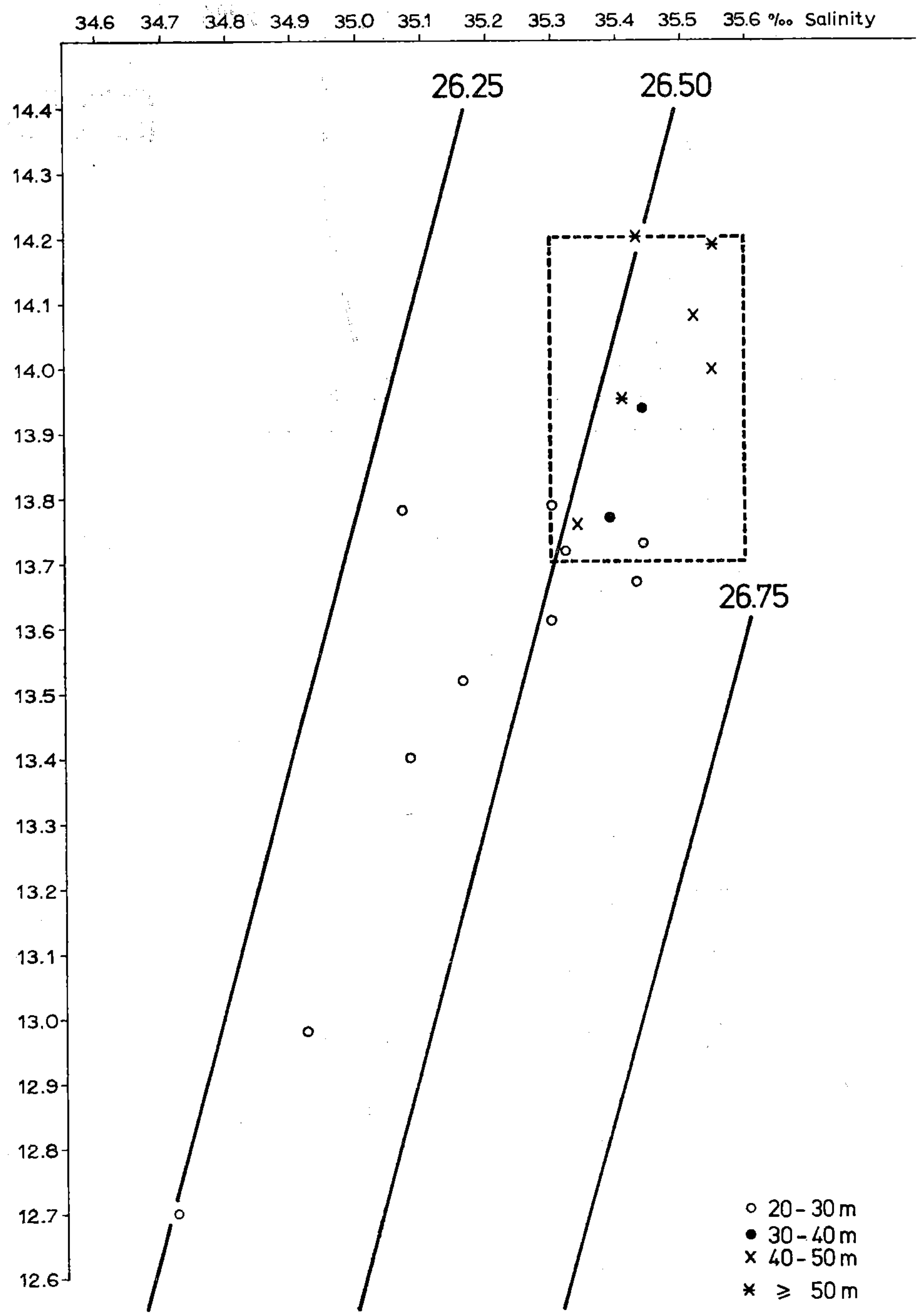


Fig. 7.6 *t-S* diagram for observations below 20 m in the Ría de Arosa (winter 1964).

atmosphere. It appears that all observations below a depth of 30 m are concentrated within a rectangle in the *t*-*S* diagram between $S = 35.3\ ^0/_{00}$ and $35.6\ ^0/_{00}$ and $t = 13.7$ °C and 14.2 °C. This area we may assume to be representative of a water mass that is present in the deeper layers and that might have come from the ocean area in front of the coast. The temperatures agree with the temperatures of the ocean 50 m deep water found at 50 m in that time of the year (see section 4.4), but the salinities are somewhat below the oceanic data; they are comparable with the surface salinities at large (section 4.3). It is therefore probable that the deep waters in the ria too are somewhat affected by freshwater.

The observations from the layer between 20 and 30 m deep already show a definite decrease of salinity and temperature. For the layers between the surface and 20 m depth the data show a further decrease, as is seen from fig. 7.7.

The properties of the brackish water in the lower courses of the Rio Ulla are not included in these diagrams, but they are shown separately in fig. 7.8. Extrapolating the line that can be drawn through the points in this *t*-*S* diagram we find a value of about 6.5 °C for the temperature of the freshwater of the Rio Ulla.

If the properties of the water in the ria were mainly caused by the mixing of river water with the deep water the observed temperature and salinity values would cluster around the mixing line between the *t*-*S* values of these two water types, which is

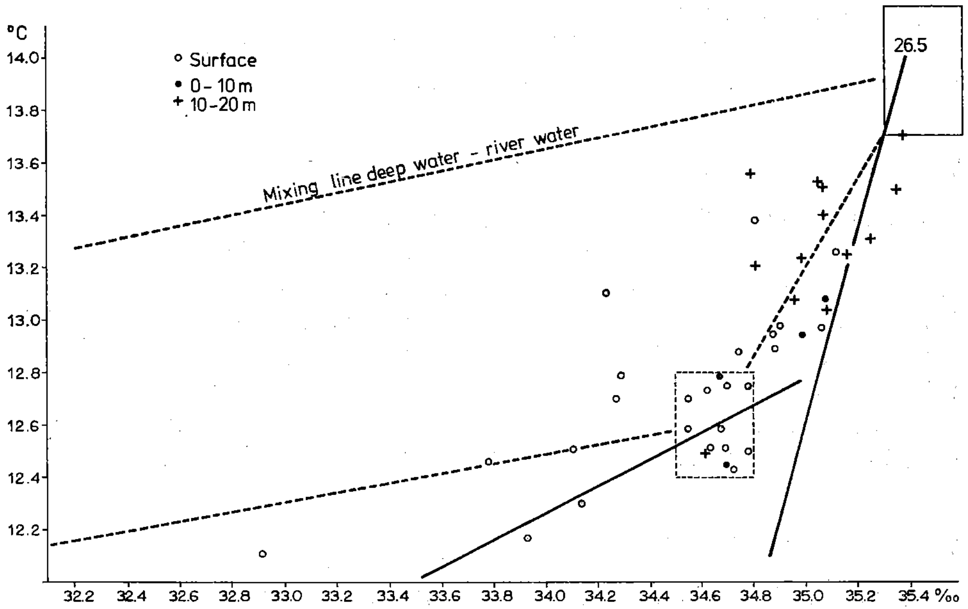


Fig. 7.7 *t*-*S* diagram for observations above 20 m in the Ría de Arosa (winter 1964).

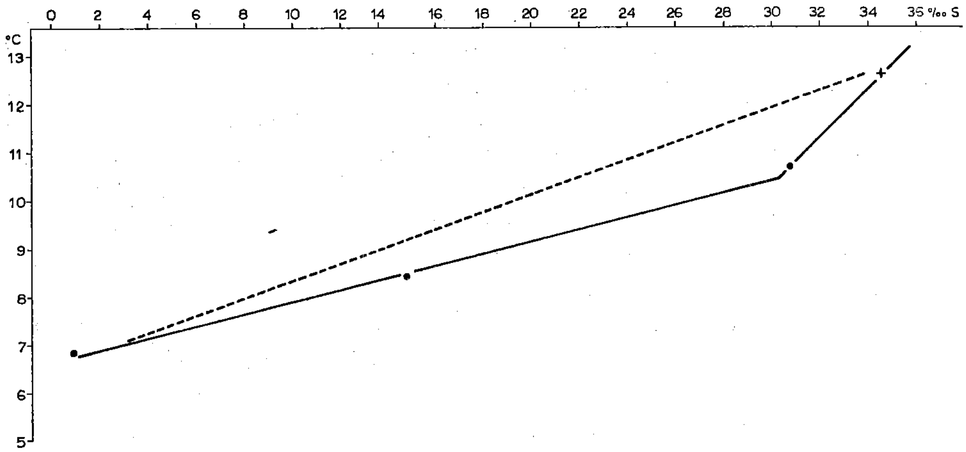


Fig. 7.8 *t-S* relation for Rio Ulla.

indicated in the diagram of fig. 7.7. In reality the observed values are far below this line, which indicates that also cooling of the water is important.

The water with salinities between $S = 34.5$ ‰ and $S = 34.8$ ‰ and with temperatures between $t = 12.4$ °C and 12.8 °C is more or less typical for the surface water, 40 % of the surface observations showing salinities and temperatures within this range. Schematically, the water of the ria can be thought to consist of a mixture of river water, the above-mentioned surface water and the deep water.

7.3 The properties of the water of the Rio Ulla

The temperature-salinity relation of the Rio Ulla water is shown in fig. 7.8. The number of observations is rather small, and therefore we can only make some general remarks. It appears from the linear relation between the t and S values that the mixing in the river itself is a rather rapid process during which the water does not cool down appreciably. The distribution of salinity along the river is such that the observed values fall within the salinity range determined from the summer observations (see fig. 6.11). The river discharge probably should be much higher than twice the summer value in order to get an appreciable freshening of parts of the estuary.

7.4 Some remarks on the circulation during the winter

No current observations were made during the winter campaign. The number of other oceanographic observations is rather small. Therefore it is difficult to investigate the circulation pattern, and any remark on this subject based on these scarce data will be more or less speculative. However, the general impression is that the pattern of the summer circulation, consisting of an outward flowing surface current and inward moving deeper water masses, also existed during this winter campaign. Only the upwelling under influence of the wind, which appeared to occur during the summer, was probably less important or absent. Possible upward movement of deep water instead may be caused by a process of entrainment of deeper water by the outflowing surface layers such as has been described by PRITCHARD (1955). The reasons why upwelling is not thought to play such an important part during winter as during summer are the following:

1. the data shown in fig. 7.7 do not indicate the presence of the deep water mass at or near the surface (above 10 m deep) in contrast to the summer data shown in fig. 6.10

2. the northerly winds are not such a regular feature in winter as during the summer.

A point which deserves consideration is the question whether the surface water cannot be cooled down so far that it attains a higher density than the bottom water. This would mean that convection would play a part in the circulation. It may be shown that the temperature decrease necessary to start such a process would be a decrease from about 12.6 °C to 10.8 °C, a decrease of about 1.8 °C (taking the surface water to have the t and S values 12.6 °C and 34.65 $^0/_{00}$, respectively). This cooling increases the density to σ_t values of 26.6, which is the density of the deeper water. Such a decrease of temperature certainly is not impossible and we may therefore assume that under certain conditions convection may play a part in the circulation. Dry and cold

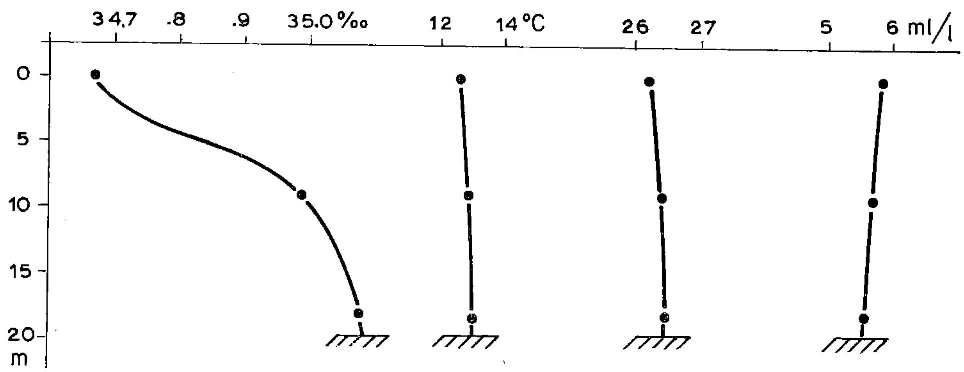


Fig. 7.9 Vertical oceanographic profile, Grove deep 3 February 1964.

weather would be favourable for convection, wet and warm weather would prevent this process.

In section 6.19 the situation in the 'Grove deep' has been discussed, especially the problem whether stagnancy of the deeper water may be expected to occur or not. The winter situation as observed on February 3 is shown in the profiles of fig. 7.9. The stability of the water-column is small, and the oxygen content near the bottom is high. A higher discharge of freshwater by the Rio Umia may increase the stability, but as also the cooling of the water in the shallow lower course of the river and in the Bay of El Grove will be very effective, it is doubtful whether this will be sufficient for a complete isolation of the water in the deep layers.

8. Turbidity and suspended matter

8.1 Introduction

The turbidity of the water is important for the oceanographic conditions of a sea area in several respects. First the photosynthesis depends on the available radiation at a certain depth, and this (among other factors, such as the length of the day, the inclination of the sun, the cloudiness and the reflection by the sea surface) also depends on the turbidity of the water. In the second place the turbidity depends on the nature and the concentration of the suspended matter in the sea, and therefore is of interest in the study of suspended particles, organic and inorganic.

An indication of the turbidity of the water is obtained by Secchi disc observations. The Secchi disc (see section 5.5) is a very simple instrument and the observations are not very accurate, depending very much on the circumstances of the observation and the ability of the observer, but they have the advantage that they are easily made in a short time. In the following we shall discuss these observations in connection with observations of suspended matter.

8.2 Secchi-depth

The observations of the Secchi-depth that were made during the successive summer campaigns were combined into a chart giving the average values, in the same way as was done for the charts of the surface salinity and the temperature. This chart is shown in fig. 8.1. The variation of the Secchi-depth within the same area is, however, considerable. The observed values range from 3 to 12 m, and small and large values occur equally over nearly the whole ria. Still there is in the average pattern a tendency towards large values in the outer ria and smaller values in the inner ria. For the Rio Ulla no data are given in fig. 8.1, but it can be added that here no value of the Secchi-depth smaller than 2 m has been observed (see section 8.6).

In the major part of the ria the average Secchi-depth lies between 5 and 7 m.

The number of Secchi-depth observations made during the winter campaign of 1964 is very small and not sufficient for general conclusions regarding the light regime. The impression is that in the inner parts of the ria the turbidity is somewhat higher in winter than in the summer. Values between 3.5 and 5.5 m were observed. In the outer parts the Secchi-depth was 6.5 to 7 m, and in the region in front of the Rio Umia remarkably high values (7 and 8 m) were found.

8.3 Maximum compensation depth

The observed Secchi-depth values may be used to find an estimate of the maximum compensation depth (see section 6.5) which according to RYTHER (1963), approximately

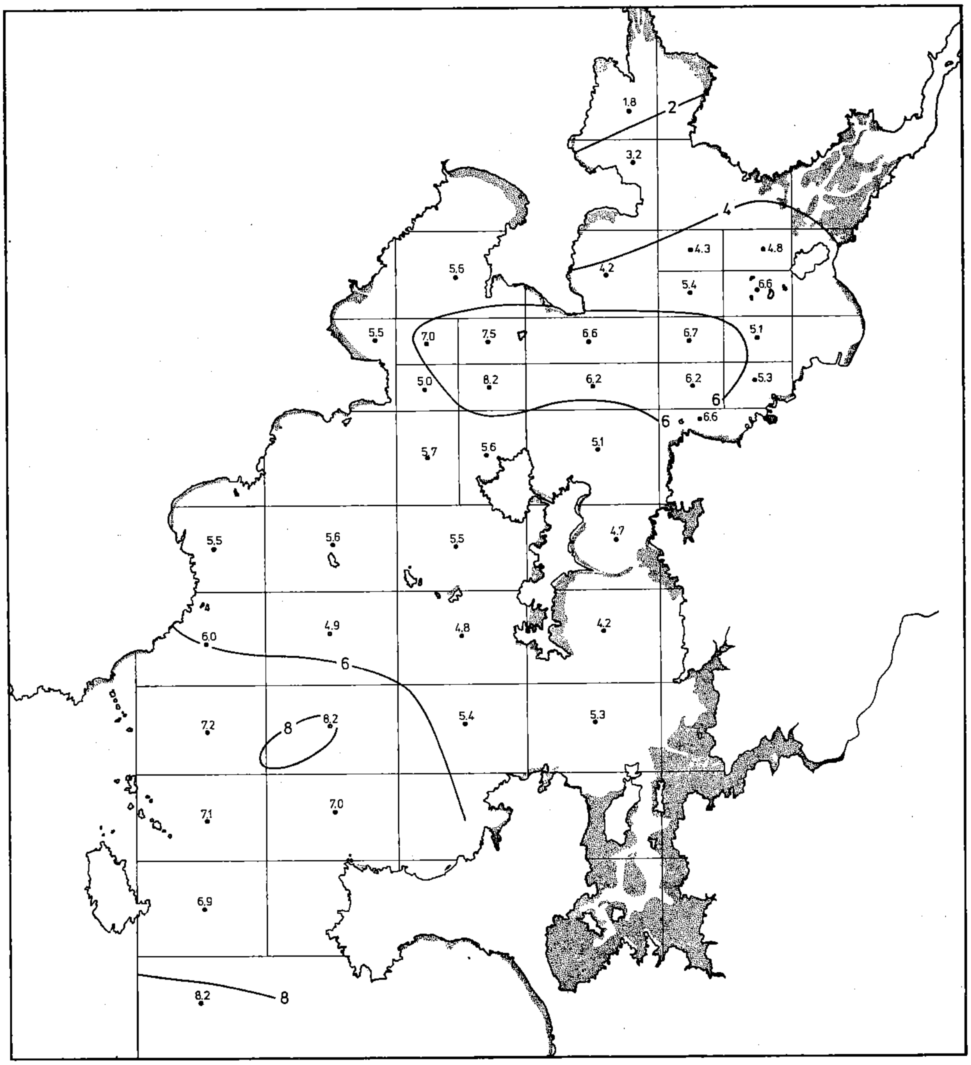


Fig. 8.1 Average secchi depth, Ría de Arosa (summer).

equals the depth to which 1 % of the incident visible radiation penetrates. At this depth oxygen production by plant growth under optimal conditions (clear sky and maximum altitude of the sun) just balances oxygen consumption. In order to find this depth we estimated the so-called 'vertical attenuation coefficient' using an empirical relation between this quantity and the Secchi-depth. Here the vertical attenuation coefficient is defined by

$$c^* = \frac{1}{z} \ln \frac{F_0}{F_z}$$

where F and F_z are the radiant flux of the daylight at the surface and at a depth z respectively. The following relation between the Secchi-depth S and c^* is more or less widely in use:

$$c^* = \gamma / S$$

where γ is a value usually between 2.5 and 1.5, depending on the circumstances under which the Secchi observation has been made (see JOSEPH, 1952). Thus for the major part of the ria the vertical attenuation coefficient is estimated to have average values between 0.2 and 0.5 m^{-1}. According to the above, the maximum compensation depth z_c is given by $e^{-c^* z_c} = 0.01$, or

$$z_c = \frac{\ln 100}{c^*}$$

It follows that in most places the maximum compensation depth on the average lies between about 10 and 25 m. In exceptionally clear water, which may occur in some places, especially in the outer ria, the Secchi-depth attains a maximum value of 12 m, which corresponds with a value of z_c between 25 and 40 m. It may be noted that, according to data from JERLOV (1961), in the ocean at some distance from the Galician coast z_c is of the order of 80 m.

8.4 Comparison with data from the Ría de Vigo

VIVES AND LÓPEZ-BENITO (1958) have presented data on the turbidity of the waters of the Ría de Vigo. These authors measured the vertical attenuation coefficient by means of an underwater photometer with a maximum sensitivity at a wave length of 520 mμ. In the upper layers the observed values varied between 0.5 m^{-1} in the inner ria in August and 0.01 m^{-1} in the outer ria in September. The stations near the mouth of the ria had a lower turbidity than the inner parts. Furthermore, there appeared to be a tendency towards high values of the attenuation coefficient in February-April and low values in July-September (with as a marked exception the high August value given above).

This seasonal variation agrees with the differences found for the Secchi-depth during our summer and winter campaigns.

The high values of the attenuation in the inner part of the Ría de Vigo have been correlated by these authors with the advection of fresh water, and their conclusion is that these high values are caused by inorganic (silt) particles. The attenuation in the

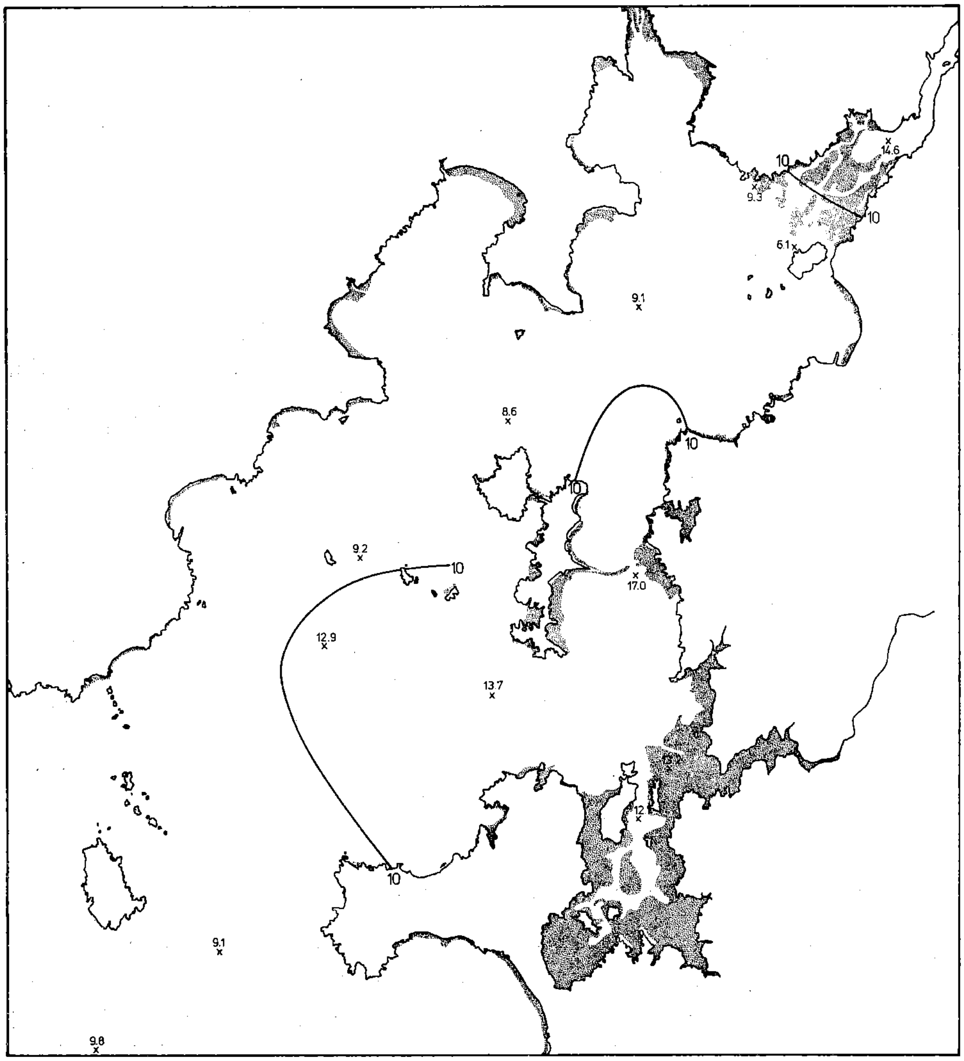


Fig. 8.2 Surface concentration of suspended matter (summer).

more oceanic parts probably largely depends on the concentration of organic particles in the water. According to FRAGA (1960) only a part of this suspension is phytoplankton, and a large part consists of other organic particles.

8.5 Suspended matter in the Ría de Arosa

The term 'suspended matter' is here used for all kinds of solid particles that may occur in sea water. These particles may be very different in nature and in their movements relative to the water. We may distinguish the following categories:

a. Small inorganic particles (silt) having a settling velocity that is so small that they are transported by currents in practically the same way as dissolved substances.
b. Larger inorganic particles (silt, sand), which have larger settling velocities than the particles mentioned under *a.* and for which their own movement relative to the water cannot be neglected.
c. Living phytoplankton, occurring mainly in the euphotic layer.
d. Living zooplankton, which may occur also below the euphotic layer and may migrate in mainly vertical direction.
e. Dead planktonic remains and detrital particles.

The observed concentrations of suspended matter in the Ría de Arosa show variations from one station to the other, but also for different depths at the same station and for repeated measurements at the same station. The different kinds of particles that contribute to the suspension and their different behaviour may give rise to variations that are often difficult to understand.

As additional information on the samples, for instance analyses of the size distribution or of the fractions of organic and inorganic particles in the suspension, is lacking, it is hardly possible to derive from the present data more than a general picture of the nature of the particles in suspension and of their distribution. This picture will largely be based on circumstantial evidence, which means that it might be improved in several respects.

8.6 Suspended matter in the rivers

The rivers that flow into the Ría de Arosa should be considered separately in this context. The shallow depths, the relatively strong currents during ebb and flood and the large gradients of the salinity may be expected to influence the character of the suspension and its concentration to a larger or smaller degree. Samples were taken in the Rio Ulla and the Rio Umia in 1962 and in 1963. The possibility that too high concentrations are found is larger for the rivers than for the surface and intermediate depth samples of the ria itself, because of the small depth of the rivers and the enlarged chance of stirring-up bottom material.

The following table gives some data on the concentration of suspended matter in the rivers.

In this table near-bottom data have been omitted, in order to minimize the effect of sampling errors. Although the possibility of such errors still cannot be excluded completely, these data are believed to indicate the occurrence of relatively elevated

TABLE X. *Suspended matter in the Rio Ulla and Rio Umia*

	Ulla	Umia
salinity $< 1^0/_{00}$ (1962)	5.1 mg/l	6.1-3.8 mg/l
$1\text{-}10^0/_{00}$ (1962)	23.0-76.4	4.6
(1963)	9.2	
$10\text{-}20^0/_{00}$ (1963)	8.4	16.6-10.3
$20\text{-}30^0/_{00}$ (1962)	6.4-10.1	5.9
$> 30^0/_{00}$ (1962)	6.0	
(1963)	14.6	

concentrations in the brackish parts of the rivers. A similar tendency is observed for Secchi disc data of the Ulla, mainly obtained in 1964 with the use of a brown disc (because the small depth does not allow the use of a normal disc). These data afterwards having been converted into normal Secchi disc values by means of an empirically obtained relationship

$$S_{\text{normal}} = 1.8\, S_{\text{brown}}$$

The data for the river Ulla are:

TABLE XI. *Secchi-disc values Rio Ulla*

salinity $< 1^0/_{00}$	$S = 3\frac{1}{2}\text{-}2\frac{1}{4}$ m	(1964)
$1\text{-}10^0/_{00}$	$2\frac{1}{4}\text{-}2$ m	(1964)
$10\text{-}20^0/_{00}$	$2\frac{1}{2}\text{-}2$ m	(1964, 1963)
$20\text{-}30^0/_{00}$	$3\frac{1}{2}$ m	(1964)
$> 30^0/_{00}$	$3\frac{1}{2}$ m	(1964)

For a suspension of a given relative size distribution a decrease of the Secchi-depth indicates an increased concentration of the suspension. Therefore these data again suggest the presence of the so-called 'turbidity maximum' in the river. The causes for such a turbidity maximum, which is found in many rivers, have been discussed by POSTMA AND KALLE (1955) and POSTMA (1967).

8.7 Concentrations of suspended matter observed in the Ría de Arosa

In the main channel of the ria observations of suspended matter have been made in 1962 during a period of about 12 hours on the anchor station mentioned in paragraph 6.13, and in 1963 at a number of oceanographic stations. The average values of the concentration of suspended matter found at the anchor station in 1962 are 7.7, 7.1 and 7.5 mg/litre at 0 m, 25 m and 50 m depth respectively. The values of the individual observations ranged between 14.3 and 4.9 mg/litre at the surface, between 8.4 and 3.6 and between 9.0 and 6.3 mg /litre at 25 and 50 m, respectively. So, especially at the surface the variations with time were rather large. The observations of the surface concentrations found in 1963 have been combined in a chart showing the distribution of the concentration of suspended matter at the surface. In view of the large variations with time observed in 1962 we should not attach too much weight to the individual values, but rather look for the general pattern. The chart given in fig. 8.2 shows a region with relatively high concentrations between Isla de Arosa and El Grove, and further to the east. Here values of 10 to 17 mg/litre were found. In the larger part of the main channel the values ranged between 9.3 and 8.6 mg/litre, with the exclusion of one station near the mouth of the river Ulla where 6.1 mg/litre was observed. We may wonder whether this distribution reflects a real feature of the concentration of suspended matter or whether it is the result of a fortuitous grouping of the values. A definite answer to this question cannot be given; the number of observations is rather small and chance might very well play a part in this case. But the point is interesting enough to be mentioned.

The values at greater depth down to a short distance from the bottom were of the same order of magnitude. We shall pay more attention to the vertical distribution of suspended matter later on, but here we shall already mention the occurrence of some rather high values near the bottom that may be the result of the sampling operation rather than a real effect. It is possible that unconsolidated sediment is disturbed by the sampling. For instance, in the inner ria at a station with a bottom depth of 12 m at a sampling depth of 10 m a concentration of 24.5 mg/litre was found. KOLDIJK (1968) has mentioned the occurrence of a 'soupy mobile layer' on the bottom in the inner ria. Such a layer is easily disturbed, especially because the bottom depth cannot always be determined accurately by sounding if such layers are present.

Concluding we may state that for the surface waters of the ria a value of 10 to 7 mg/litre may be considered representative for the main channel, but that temporal and regional variations may be expected, resulting in maximum values up to 17 mg/litre and minimum values that may be as small as about 4 mg/litre. These values are several times the concentrations in the surface waters along the Spanish and Portuguese coasts as reported by SVIRENKO (1970), who in summer 1966 found concentrations between 0.6 and 1.4 mg/litre (average 0.9 mg/litre).

8.8 Nature of the suspended particles

Various factors may give indications of the nature of the suspended particles. These are the distribution of the concentrations and the Secchi-depth. If we have a suspension of particles with a constant relative size distribution there will be an approximately linear relation between the inverse Secchi-depth and the concentration. Otherwise the relation will be more complex, and a certain concentration (by weight) of small particles will give a smaller Secchi-depth than the same concentration of larger particles, if their other properties are the same. POSTMA (1961) has used the relation between Secchi-depth and concentration of suspended matter for the determination of the so-called 'optical diameter'.

If the suspension consists exclusively of small inorganic particles from sea or river, which settle down so slowly that they are transported in practically the same way as dissolved substances, one may expect a linear relation between salinity and concentration of suspended matter and (as long as the size distribution does not change appreciably) a similar relation between salinity and the inverse Secchi-depth. From the observations of 1962 and 1963 (only surface and intermediate-depth data) diagrams were made of the concentration of suspended matter versus salinity and of the inverse Secchi-depth versus the salinity at the surface.

No clear linear relation is found between concentration of suspended matter or Secchi-depth and the salinity. For the 1962 data a linear relation might exist of the extremely high concentration of 14.3 mg/litre could be neglected and for the 1963 data a linear relation can be found for the data with salinities below about 33 $^0/_{00}$. But more data are needed before a definite conclusion can be drawn. For the moment our conclusion can only be that at least for the major part of the ria there is a considerable contribution of other than silt particles to the suspension. This agrees with the findings for the Ría de Vigo (see section 8.4).

Looking at the vertical distribution of suspended particles in intermediate waters, from 20 m below the surface down to 10 m from the bottom, we find values ranging between 8.9 and 10.5 mg/litre, with an average of 9.4 mg/litre. The observations made in the interval from the bottom up to 10 m range from 11.3 to 26.4 mg/litre, with an average of 17.5 mg/litre. This includes data from regions where the sediment is not as easily disturbed as in the inner ria. The 1962 observations did not show such a vertical distribution, the near-bottom values of the concentration being only slightly higher than those at the surface or at intermediate depths. Our conclusion is therefore that the concentrations tend to increase towards the bottom, but that this increase may be very different at different places and different times.

As regards the rate of sedimentation we may refer to Pannekoek (in HINZ, 1970), who estimated a rate of sedimentation of about 1 m in 1000 years. According to CADÉE (1968) the organic carbon content of the sediment is between 1.6 % and 7.0 %. Even if this organic material would settle down exclusively in the summer months the amount of organic carbon withdrawn from the suspension by sedimentation would be of the order of 2 mg per month or less for an area of 1 cm^2. At a depth of 10 m this means 2 mg/litre per month. Compared with the quantities of particles in suspension, which are supposed to be for a large part of organic nature, this means a rather low rate of withdrawal. Therefore for the balance of suspended organic carbon, dealt with in section 10.10, it is thought that sedimentation plays no significant role.

9. Chemical data

9.1 Introduction

Observations of various chemical properties of the water in the ria have been made during the different campaigns, as indicated in the following table:

dissolved oxygen	all campaigns
pH values	summer 1963 and summer 1964
nitrate and nitrite	winter and summer 1964
phosphate	winter and summer 1964
dissolved organic carbon, summer 1962 and summer 1963	only a small number of samples
dissolved organic nitrogen and ammonia, winter 1964	

The oxygen, phosphorus and nitrogen data show some deviations from what might be expected. From many investigations in other places it has appeared that in a certain water mass, especially below the euphotic layer, the concentrations of inorganic phosphorus and nitrogen usually show a certain negative correlation with the concentration of dissolved oxygen. These interrelations have been discussed by several authors, for instance by REDFIELD, KETCHUM AND RICHARDS (1963). They are the result of oxygen consumption during decomposition of organic matter, by which process phosphate and nitrate are produced, and of the reverse process of photosynthesis. Accordingly, the following relation was found to hold approximately for the variations of atomic concentrations of the elements concerned:

$$\triangle O : \triangle N : \triangle P = -276 : 16 : 1$$

(REDFIELD, KETCHUM AND RICHARDS, 1963)

When it was investigated whether the ratios between the concentrations of dissolved oxygen, of phosphate-phosphorus (PO_4-P) and nitrate- and nitrite-nitrogen (NO_3-N plus NO_2-N) are in accordance with this rule, negative results were obtained. This was not only so for the whole collection of data, but also for selected groups representing one water mass. This does not, by itself, mean that the observed values are erroneous. We should note that, according to ARMSTRONG (1965), especially in enclosed waters the rule of a constant nitrogen-phosphorus ratio has its exceptions. This is not the place to go further into this subject. However, it leaves us without an objective method for checking the individual data. This is especially regrettable, as there is a need for such a form of quality control. As the conditions under which the work had to be done were rather primitive, we are not completely certain that each individual observation is as accurate as could be expected from the analytical technique alone. But things being as they are, we may note that the observed values are in any case comparable with similar data from the Ría de Vigo. Therefore we have confidence that the observed values can be used in a general discussion and that the accuracy of average values will be acceptable, but we believe it unrealistic to deal with individual observations in detail.

A presentation of some of the oxygen data has already been given in connection with temperature and salinity data (see Chapter 6). As shown in fig. 6.4 and 6.17 concentrations below 4 ml/l are observed at greater depths. There is a tendency for relatively low values to occur at the righthand side of the estuary. Such a phenomenon, if real, can be due to either longer residence times of the water at greater depths, reduced mixing (because of greater vertical stability) or higher oxygen consumption (because of a higher input of organic material).

9.2 Variations of oxygen content at Punta Preguntoiro

The diurnal variation in the relative importance of photosynthesis and decomposition of organic matter may be estimated from the variations of the concentration of dissolved oxygen. These variations may, however, be caused not only by these processes but also by advection of water masses with different oxygen content or by exchange with the atmosphere at the surface.

Regular oxygen observations over a period of 24 hours have been made at two stations: on 11-12 July 1963 at Punta Preguntoiro and on 19-20 July 1963 near La Toja. Besides further data from a 24-hours station made by the 'Xauen' of the Instituto Español de Oceanografía in September 1950 may be considered (INST. ESP. OCEANOGR., 1955). However, only at the first station the situation is such that the effect of advection may be considered to be of minor importance. This is confirmed by the salinity and temperature data.

The results of these observations are shown in fig. 9.1. The salinity shows only small and irregular variations until the afternoon of the 12th July, when a rapid decrease of the salinity, possibly associated with a rain-shower that had started a little earlier, was observed. The observations were made at the end of the landing pier, in rather shallow water with rocks covered by an algal vegetation (locality 7 in DONZE, 1968). This location therefore is not representative for the ria as a whole, but the observations are nevertheless considered to be of interest. The general trend is that during the dark hours the oxygen concentration is rather low, the water being slightly undersaturated (about 95 %). After sunrise the oxygen concentration increases and in the afternoon values around 125 % are observed. After sunset there is again a decrease of oxygen concentration.

If we assume that the exchange with the atmosphere as well as the effect of the advection of different water masses is negligible, we conclude that there has been a production of (at most) about 1.5 ml/l oxygen, which would mean a production of

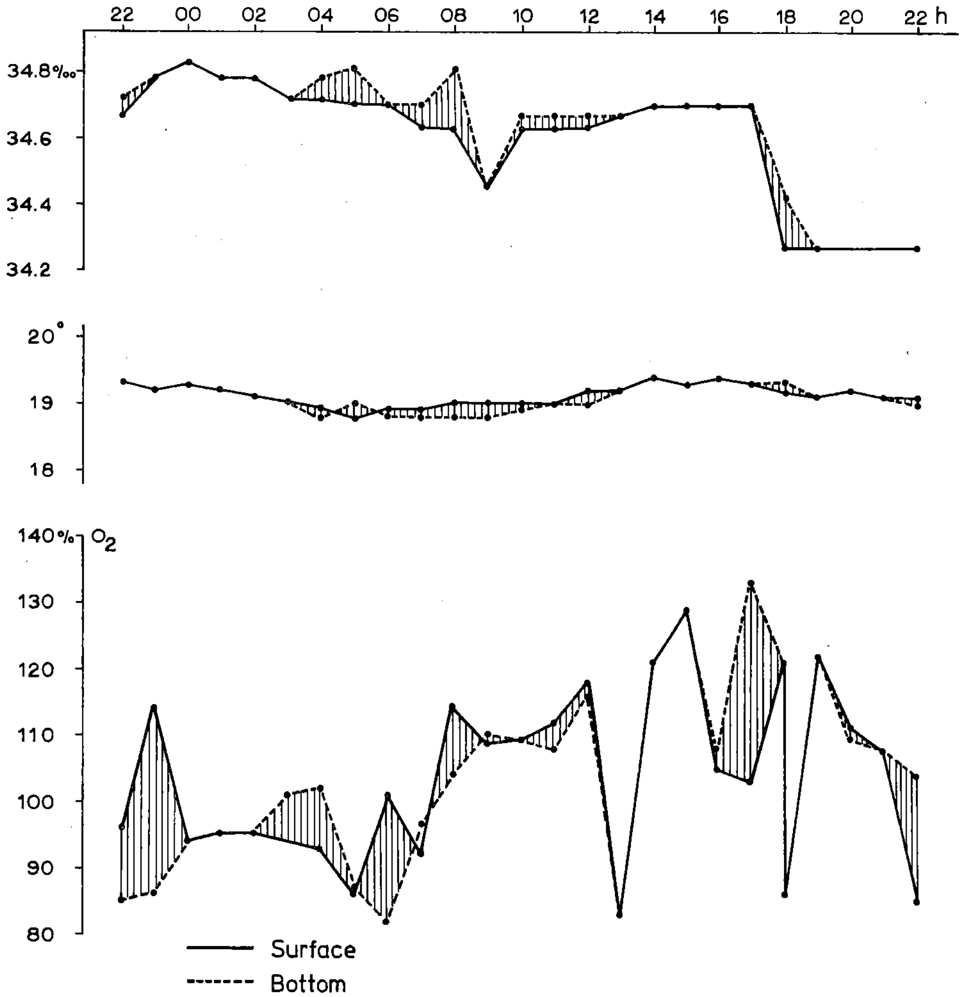


Fig. 9.1 Diurnal variation of oceanographic properties at Punta Preguntoiro, 12 July 1963.

about 50 µg-atoms or 0.6 mg organic carbon per litre. This value can be regarded as the mean of a layer of at least 3 m thickness (the maximum distance between the surface and the bottom observations), which gives a production of roughly 2 g carbon/m^2 day. As loss to the atmosphere is not taken into account, this value is a minimum value.

Although this location cannot be regarded as fully representative, the result is in rather good agreement with the value given by STRICKLAND (1965) for the production in coastal upwelling areas in the trade-wind regions, i.e. of the order of 1.5-2 g carbon/ m^2 day, and with the estimated production values given in section 10.9.

9.3 Oxygen saturation at the surface

Although the oxygen concentration in the near-surface layers may exhibit considerable variations with time, we may assume that during daytime (when most of the

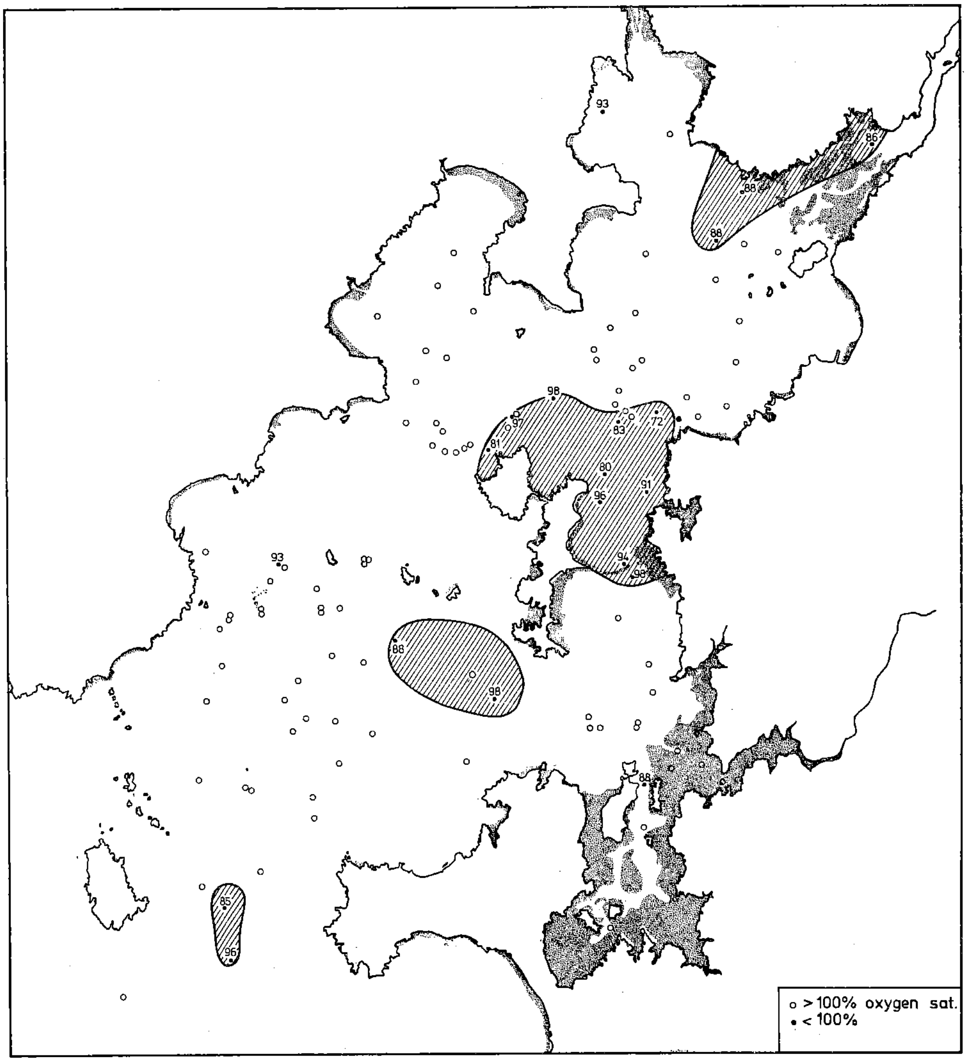


Fig. 9.2 Mean oxygen saturation surface water, Ría de Arosa, summer

observations were made) the observations from various places may be well comparable. Because of the photosynthetic activity of the phytoplankton the surface values will then be at a maximum and a certain degree of supersaturation may be expected. Only in cases where decomposition of organic matter exceeds photosynthesis, or where water is rapidly cooled (by which the solubility of oxygen is increased), or where undersaturated water from greater depth has recently ascended or has been mixed with the surface waters, undersaturation may occur. In our case, where the water at the surface is warmed-up, only the first and the last possibility may apply.

In fig. 9.2 the positions are indicated where undersaturated water was observed at the surface. Apart from some isolated observations the majority of the observations is found in the inner part of the ria, and especially in more or less coherent groups north of the Isla de Arosa and near the mouth of the river Ulla.

These observations suggest the presence of decomposing organic matter or the occurrence of upwelling in these regions, or a combination of these two effects. The low water temperatures and the current pattern discussed in the previous chapters gave strong indications for upwelling to occur in the inner parts of the ria, and this is confirmed by the present data. However, the undersaturation near the Ulla may to some extent also be caused by advection of organic matter by this river.

9.4 Nutrients in the Rio Ulla and Rio Umia

Before considering the distribution of nutrients in the Ría de Arosa we shall discuss the concentrations in the river water flowing into the ria.

In fig. 9.3 and 9.4 the results of simultaneous observations of salinity and of phosphate and nitrate concentrations are shown for the two rivers. A straight line has been drawn in these figures between the points with minimum salinity and with the average salinity of the ria surface water (see the next section). This straight line indicates the points in the diagram that would be found if mixing were the only cause of the variations of the nutrient concentrations along the estuary.

As may be seen the deviations from this line are much greater for the Rio Umia than for the Rio Ulla. Apparently, in the Umia other processes than mixing are also of importance. The generally negative deviations suggest that within the Umia estuary part of the nutrients is already consumed by photosynthesis before entering the ria. It is interesting to note that the pH values of brackish water of the Umia were found to be much higher than those of the Ulla (8.5-8.7 against 7.7-7.9), which may also be explained by a higher photosynthetic activity in the Umia than in the Ulla. It may further be mentioned that the temperature-salinity relation for the Umia also turned out to be rather irregular (section 6.4), which could be explained by the special topography of the river mouth. During low tide the lower course of the river consists of a number of more or less isolated pools, in which conditions may prevail that might be responsible for a relatively high productivity.

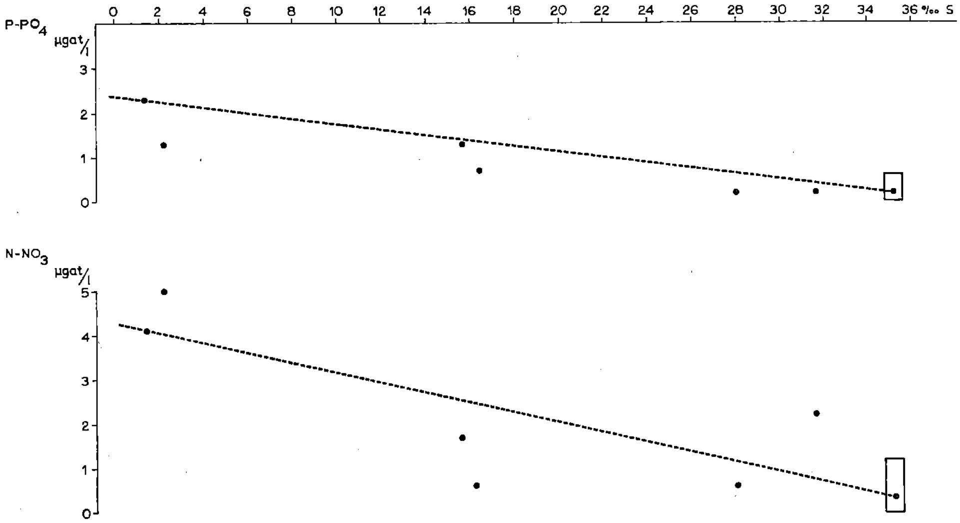


Fig. 9.3 Nutrient concentrations observed along Rio Ulla

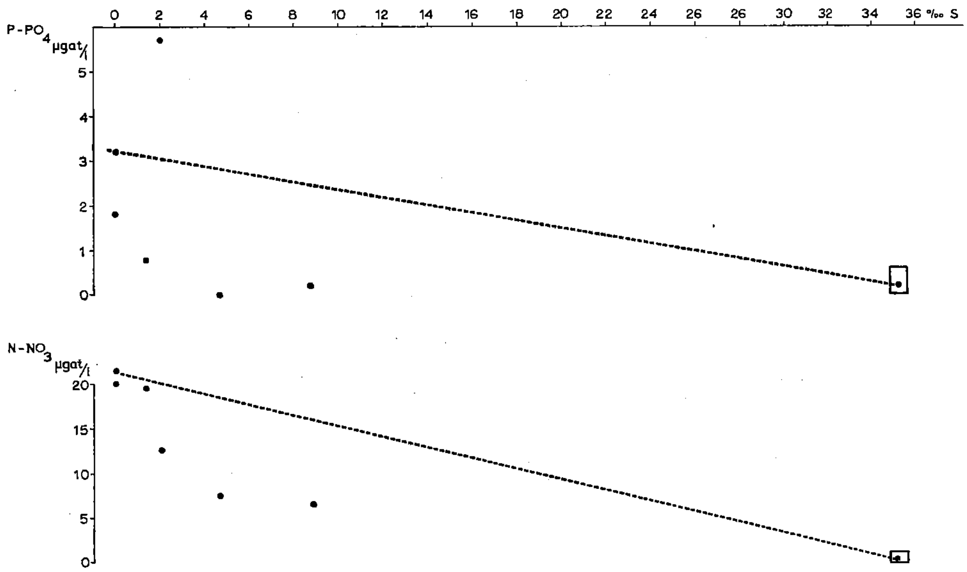


Fig. 9.4 Nutrient concentrations observed along Rio Umia.

The observations made during the winter of 1964 are scanty and do not permit any conclusions. Only some observations were made in the Rio Ulla where a constant PO_4-P concentration of 0.7 µg-at/l was observed, while the NO_3-N concentration ranged from about 17 µg-at/l in nearly freshwater via about 4 µg-at/l at $S = 15$ ‰ to the normal ria surface values (about 1.5 µg-at/l).

9.5 Nutrients in the Ría de Arosa

During summer we may distinguish several water masses in the ria (see section 6.3). In the surface water mass described as 'ria surface water', at depths of less than 10 m, and in the North Atlantic Central Water and the 'high salinity water' below 10 m depth the mean nutrient concentrations are as given in Table XII.

TABLE XII. *Nutrient data for different water masses of the Ría de Arosa (summer)*

Ria surface water,	< 10 m depth	0.2 µg-at/l PO_4-P	(range: 0- 0.5 µg-at/l)
		0.3 µg-at/l NO_3-N	(range: 0- 1.2 µg-at/l)
N.A.C. water,	> 10 m depth	1.0 µg-at/l PO_4-P	(range: 0.4- 1.4 µg-at/l)
		6.6 µg-at/l NO_3-N	(range: 0.9-10.0 µg-at/l)
High sal. water,	> 10 m depth	1.1 µg-at/l PO_4-P	(range: 0.5- 1.6 µg-at/l)
		7.8 µg-at/l NO_3-N	(range: 5.1-12.5 µg-at/l)

There is no significant difference between the two deep water masses, but they show a marked difference with the surface water, which is considered to be the result of photosynthesis at the surface and decomposition of organic matter in the deep layers. It should be noted that the phosphorus values of the deep water are higher than those observed at sea (fig. 4.8). As far as can be judged the phosphate and nitrate values are comparable with the values of the Ría de Vigo (MASAGUER AND LOPEZ-BENITO, 1967; FRAGA, 1967).

As regards the concentration of nitrite the general tendency is that at the surface low values occur (on the average 0.1 µg-at/l) with a range from 0 to 0.4 µg-at/l). Below 10 m depth the concentrations are higher (average value 0.4 µg-at/l with a range from 0 to 0.8 µg-at/l). No significant variation with depth can be observed below 10 m.

For the observations made during the winter of 1964 three water masses could be distinguished: river water, deep water and surface water, probably consisting of a mixture of the other two, cooled down by about one degree (section 7.2). The nutrient data from below 30 m depth, all belonging to the deep water mass and those of the surface water (all from the layer of 0-10 m depth) are as given in Table XIII.

TABEL XIII. *Nutrient data for the Ría de Arosa, winter 1964*

Deep water,	> 30 m depth	0.4 µg-at/l PO_4-P	(range: 0.2-0.8 µg-at/l)
		2.6 µg-at/l NO_3-N	(range: 2.1-3.0 µg-at/l)
Surface water,	< 10 m depth	0.3 µg-at/l PO_4-P	(range: 0.2-0.5 µg-at/l)
		1.5 µg-at/l NO_3-N	(range: 1.3-1.6 µg-at/l)

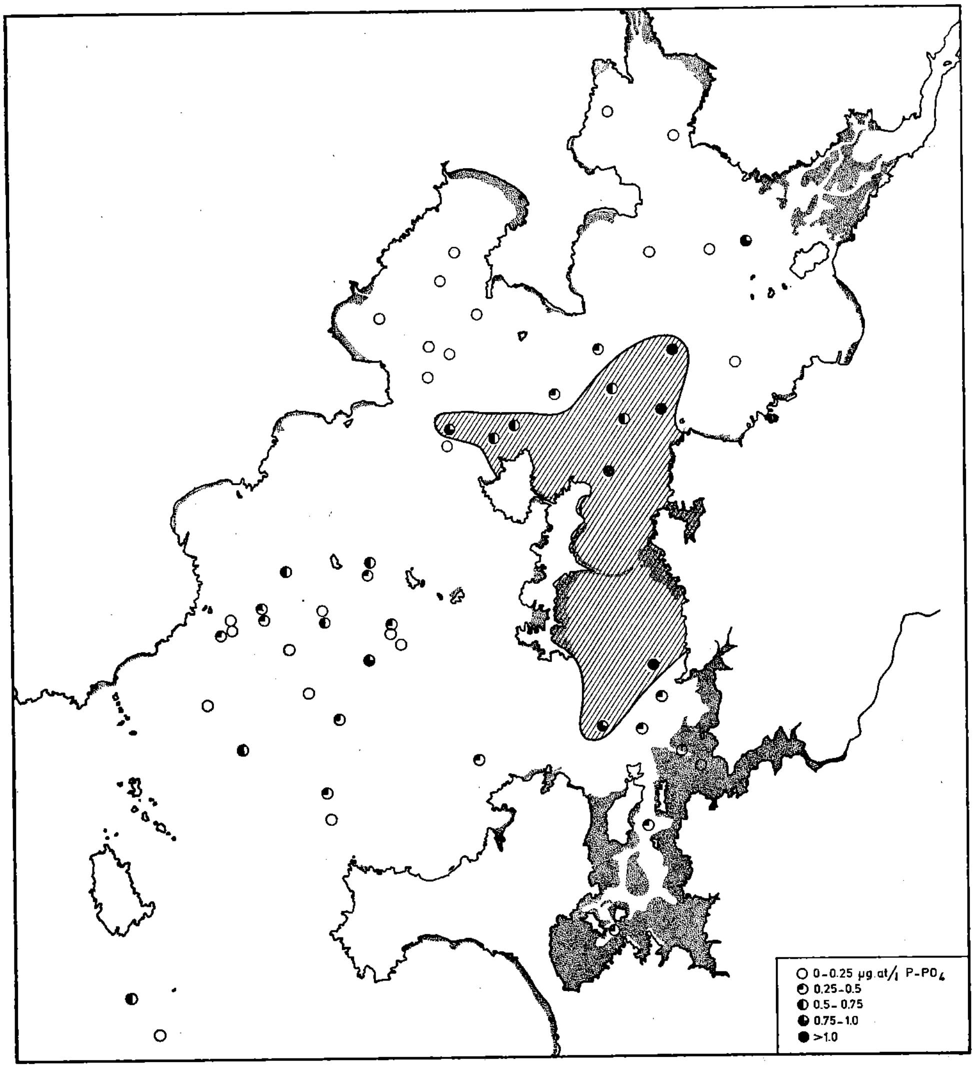


Fig. 9.5 Distribution of phosphate-phosphorus at the surface, summer 1964.

In the surface waters the concentration of nitrite is less than 0.1 µg-at/l, while at greater depths the average concentration is 0.2 µg-at/l, the ranges being 0-0.2 and 0-0.8 µg-at/l, respectively.

As regards the regional variations of the nutrient concentrations we may state that these variations are rather irregular. It is, however, interesting to consider the distribu-

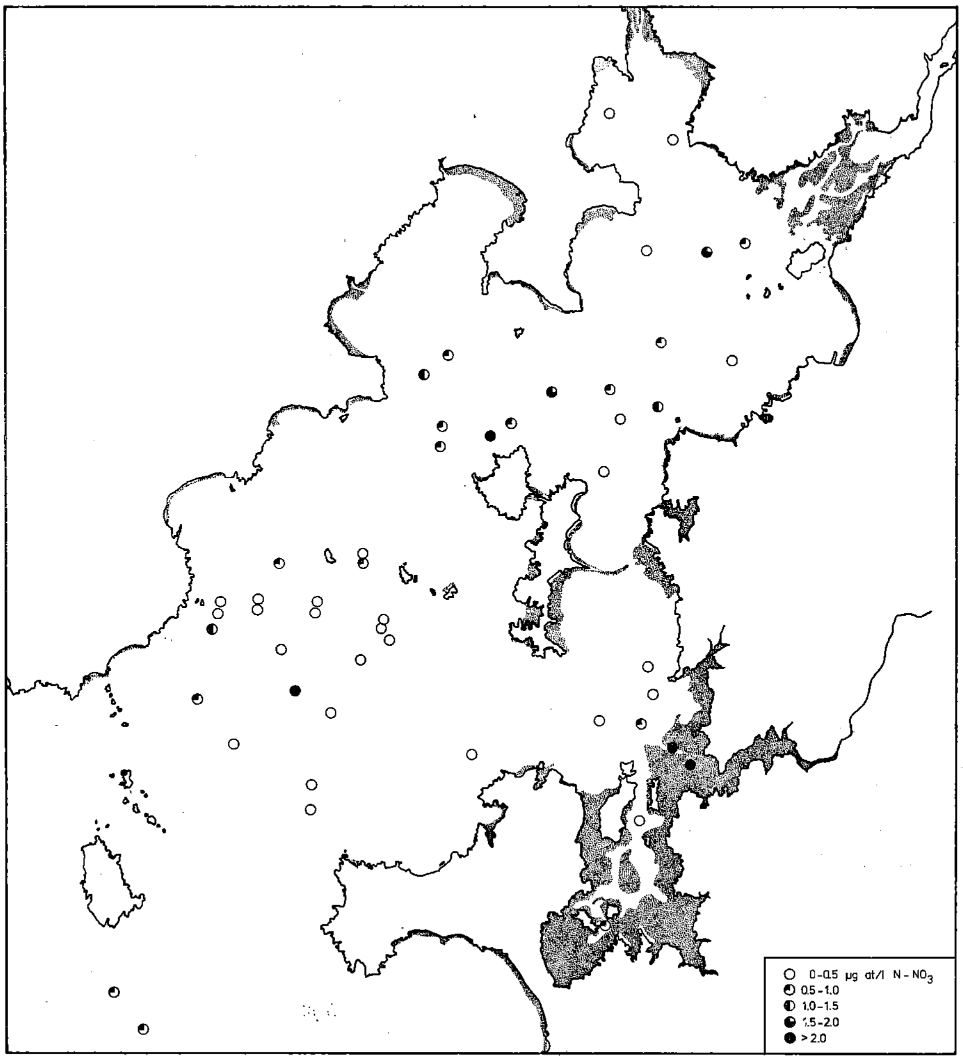


Fig. 9.6 Distribution of nitrate-nitrogen at the surface, summer 1964.

tion at the surface. In fig. 9.5 the results of surface observations of phosphate-phosphorus in summer are presented. There is a clear tendency towards elevated values in the same area of the inner ria where also low oxygen values and low temperatures are observed. This agrees with the supposed upwelling of deep water in this region. The distribution of nitrate-nitrogen generally shows low values at the surface and the few observations with higher concentrations are more scattered than the phosphate values, but still they also show some preference to occur in the inner ria (see fig. 9.6).

9.6 Observations of dissolved organic carbon

In 1962 and 1963 a small number of samples was taken for analysis of the concentration of dissolved organic carbon. The number of observations is too small to get anything but a general impression of the order of magnitude. Figure 9.7 shows the stations for which the concentrations were determined. The observed values are given in Table XIV.

TABLE XIV. *Some data on the dissolved organic carbon in the Ria de Arosa*

Station	depth	concentration of diss. org. carbon
62/222	0 m	0.82 mg/kg
	20	0.49
	40	0.56
	60	0.86
63/55	1	1.09
	20	1.01
	60	0.94
63/58	0	1.40
	20	0.85
	50	0.76
63/60	0	1.26
	10	1.02

These figures show a decrease with increasing depth (not considering the 60 m value close to the bottom of station 62/222).

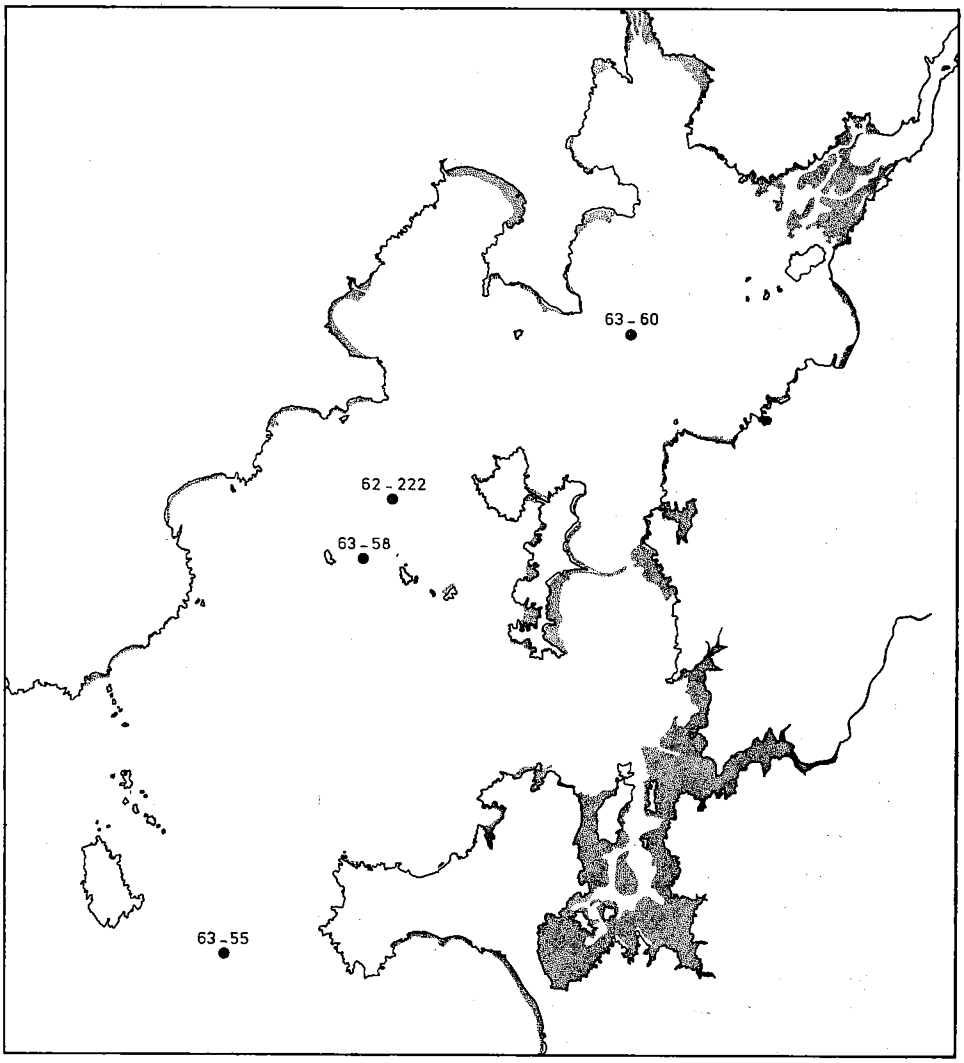


Fig. 9.7 Positions of observations of dissolved organic carbon.

10. Circulation and mixing in the Ría de Arosa

10.1 Introduction

In the previous chapters the various observations have been discussed. For the summer a circulation pattern emerges that as its main characteristic has a two-layer circulation, apparently influenced by the prevailing wind. As the northerly winds, at least in the inner ria, are the most frequent and also the mean wind direction and mean wind-stress are from the north, the circulation found for northerly winds may be considered to be the normal condition.

Combining all our observations into one picture we may draw up a circulation pattern as sketched in fig. 10.1. Although this pattern in part is based on indirect evidence (data of temperature, salinity, oxygen and nutrients) we may have confidence in its general outline. But it may be that in certain years the duration and intensity of this circulation are different, as is suggested by the difference between our data and those of GÓMEZ GALLEGO (1971).

In this chapter we aim at developing a simple model that can be used in the further interpretation of the observed distribution of the various physical and chemical properties. This is coupled with a discussion of the dynamics of the circulation, with special emphasis on the wind effects.

The circulation and mixing occur in different ways in different parts of the whole system.

In the mouth of the Río Ulla as far upstream as the sea water penetrates, there are vertical differences in salinity, and differential movement of water in different layers combined with strong vertical mixing probably is the most important factor in the upstream penetration of sea water. This circulation pattern belongs to the class 3 estuary of BOWDEN (1967) or the type B estuary as described by PRITCHARD (1955) (see section 11.1).

In the inner and central ria the circulation is characterized by a two-layer system, with upward movement of water from the lower into the upper layer by upwelling or perhaps also by such processes as have been described by PRITCHARD (1952, 1955), which occur when internal waves at the interface between both layers become unstable and break. The driving forces of the two-layered circulation in the ria may be the wind, the density distribution or both factors combined. An evaluation of these two factors will be given in the next section. Apart from this the tide will play a part in the mixing process.

In the outer ria where the greater width favours lateral variations and where generally smaller wind velocities occur, inflow occurs not only at the deeper levels, but also along the left side at the surface. Lateral mixing and transverse currents may contribute to the mixing process.

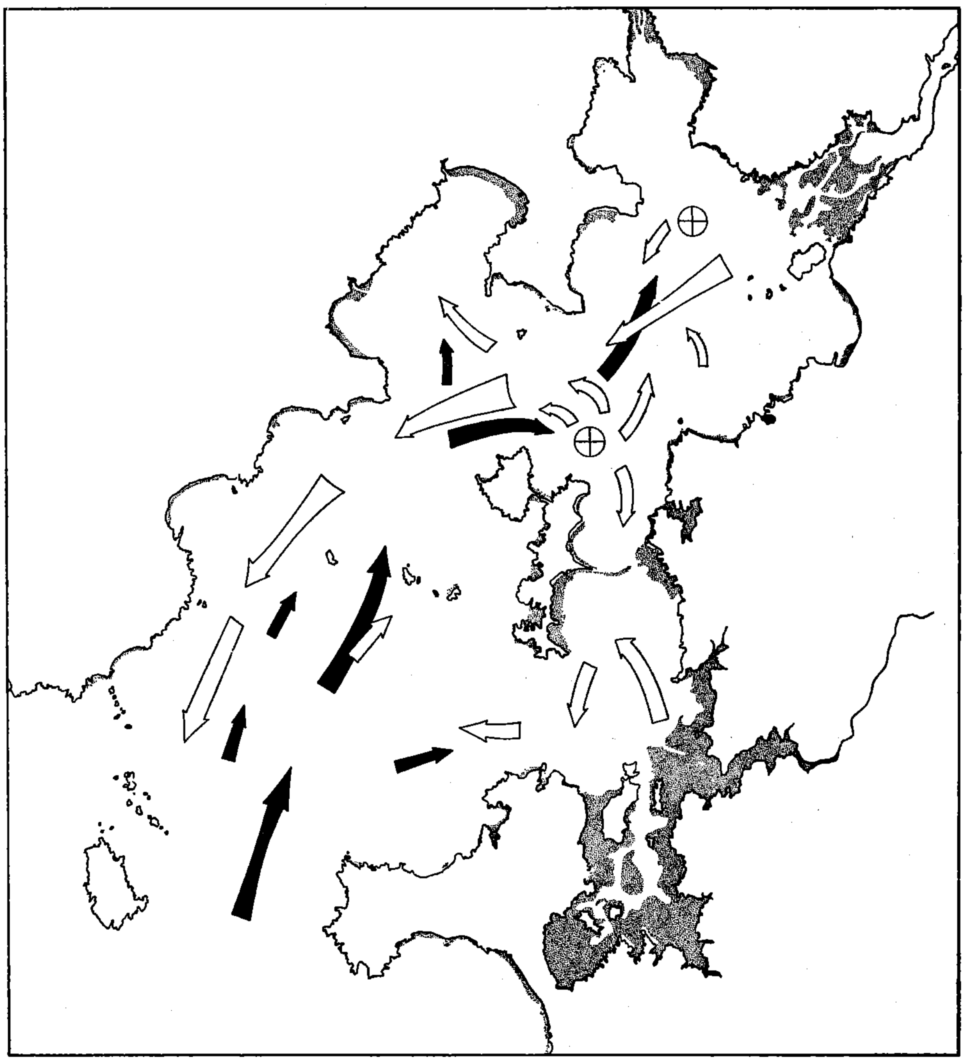


Fig. 10.1 Inferred residual circulation pattern. White arrows: surface current. Black arrows: subsurface current. Circles: upwelling.

The winter situation observed apparently was not characteristic for a normal winter and the number of observations was small. It may be that convection contributes significantly to the mixing process, but at the moment this is only a conjecture. Therefore in a further discussion of the circulation and mixing we shall direct our attention to the summer situation only.

10.2 Wind effect

For an investigation of the wind effect we not only have to know the topography of the area under investigation and the surface wind stress, but also the bottom roughness and the eddy viscosity of the water.

The topography of the Ría de Arosa is rather complicated. We may, however, consider an approximation of it in the form of a rectangular bay with dimensions comparable with those of the ria.

The wind stress at the sea surface is given by the relation

$$\tau_s = C\rho_a W_z^2$$

where ρ_a is the density of the air (about 1.25×10^{-3} g/cm^3), W_z is the wind speed at a distance z above the sea surface and C is the so-called drag coefficient. This drag coefficient depends on the level z at which the wind is measured, on the wind speed (at least accoring to some authors), and on the stability of the air above the sea surface. Different values for C suggested by several authors not only reflect the physically different conditions for which C is determined, but also the various methods used for estimating this coefficient, each of them having its special difficulties and defects.

In a review on this subject, DEACON AND WEBB (1962) give the following relation for C (at a level of 10 m) under neutral conditions of the atmosphere (W in m/s):

$$C_{10} = (1.00 + 0.07\,W_{10}) \,.\, 10^{-3}$$

Elsewhere, DEACON (1962) gives a different relation, valid for wind speeds below 14 m/sec:

$$C_{10} = (1.10 + 0.04\,W_{10}) \,.\, 10^{-3}$$

In summer stable conditions generally prevail in the ria, which means that C will be lower than under neutral conditions, so the relations given above tend to overestimate slightly the value of C.

The most frequently occurring winds in summer are the northerly and northeasterly winds, with a wind force of 4 to 5 Bft. (see table A IV), that is with wind speeds of 5.5 to 10 m/s. For these wind speeds we find an upper estimate of C_{10} by using the first of the two relations given above. (The difference is only small.) Applying this relation we find for the most frequently occurring surface stress

$$0.05 \leqslant \tau_s \leqslant 0.21 \text{ kg/m} \,.\, \text{s}^2$$

The mean stress calculated from the 8 and 20 h frequency distributions (vectorial mean) is about 0.05 kg/m . s^2 to the south. In our computations we shall therefore use a wind stress of 0.05 kg/m . s^2 to the south as a value representative for the average and most frequent situation.

The bottom friction stress τ_b may be calculated from the current velocity close to the bottom by means of the relation

$$\tau_b = k\,\rho_w u^2$$

where ρ_w is the water density and k is the bottom friction drag coefficient which depends on the roughness of the bottom and the distance from the bottom at which the current velocity u is taken. The drag coefficient is constant for neutral stability of the water mass and hydrodynamically rough flow, which is assumed to occur at high values of the Reynolds number (STERNBERG, 1970). It is assumed that under these conditions the so-called laminar sublayer is completely disrupted by the larger roughness elements. For lower values of the Reynolds number ('smooth' flow or 'transitional' flow) the value of k depends on the current velocity u.

As the current is composed of a tidal current varying as a cosine function with an amplitude u_1 and a residual current u_0, the bottom stress contains periodic and non-periodic components. BOWDEN (1956) has derived approximations of τ_{b0} for different values of the ratio u_1/u_0 (assuming the values of k to be constant). The non-periodic component τ_{b0} plays a part in the equations of motion for the tidal-average wind effect. According to Bowden we may write

$$\tau_{b0} = f k\,\rho_w u_1 u_0$$

where the factor f depends on the ratio u_1/u_0.

Now, the current velocity close to the bottom is not known, but from the data given in section 7.11 it may be assumed that $u_1/u_0 > 1$ in which case

$$f = \frac{1}{\pi}\left[\left(\frac{2u_0}{u_1} + \frac{u_1}{u_0}\right)\arcsin\frac{u_0}{u_1} + 3\left(1 - \frac{u_0^2}{u_1^2}\right)^{\frac{1}{2}}\right]$$

This expression may further be simplified for $u_1/u_0 \gg 1$, as is the case in many coastal waters, but here we use the full expression as the amplitude of the tidal current is not so large.

As regards the drag coefficient for a rough bottom we find different values in literature, depending on bottom conditions or the so-called bottom roughness parameter. If these values are adapted to the current velocity at 1 m above the bottom we find drag coefficients as summarized in the following table.

TABLE XV. *Literature data on bottom drag coefficient*

Place	Sediment, bottom configuration	Range of current speed	k	Author
San Diego harbour	—	15-26 cm/s	10^{-2}	Revelle and Fleming (in SVERDRUP ET AL. 1942)
Alverstrømen	—	up to 30 cm/s	8.8×10^{-3}	Mosby (in STERNBERG 1966)
Red Wharf Bay	sand	25-40 cm/s	3.4 4.7×10^{-3}	CHARNOCK (1959)
Red Wharf Bay	fine sand with some shingle	up to 50 cm/s	3.5×10^{-3}	BOWDEN, FAIRBAIRN AND HUGHES (1959)
?	gravel-sand	15-23 cm/s	3.6×10^{-3}	LESSER (1951)
?	mud-sand	27-40 cm/s	3.9×10^{-3}	id.
Juan de Fuca Strait	rock-gravel	up to 45 cm/s	2.2×10^{-3}	STERNBERG (1968)
Puget Sound	shell-gravel	up to 47 cm/s	3.9×10^{-3}	id.
id.	medium sand	up to 30 cm/s	2.6×10^{-3}	id.
id.	small ripples (4-6 cm high)	up to 41 cm/s	2.3×10^{-3}	
Juan de Fuca Strait	fine sand	up to 36½ cm/s	4×10^{-3}	id.

STERNBERG (1970) gives values for the drag coefficient for 'transitional' flow, observed near the deep-sea bed with different roughness elements estimated to be 0.25-1 cm high, and with velocities ranging from 0.5 to 8.5 cm/s. There is a wide variation of $\boldsymbol{k}$ under these conditions, from about 1 to 17 $\times$ 10^{-3}.

Information on the nature of the bottom of the ria is given by KOLDIJK (1968). As the bottom conditions are different for various regions of the ria, it is difficult to give a single value for $\boldsymbol{k}$. In fact the conditions in the ria may encompass all situations presented in the table given above. Moreover there is some correlation between the nature of the bottom and water depth. Especially in the shallower part of the ria, with a very soft muddy bottom it is even possible that the flow is 'transitional' to smooth.

An upper estimate for the value of the bottom stress may be attempted. The amplitude of the average tidal current $\bar{u}_1$ was estimated to be between 20 cm/s and 5 cm/s (see section 6.9). The rather wide range of $\bar{u}_1$ makes an estimate of the value of u_1 at 1 m above the bottom an unnecessary refinement. We shall therefore give calculations of τ_{b0} for tidal amplitudes at 1 m above the bottom of 20 cm/s.

The estimates of the residual current given in section 6.11 may be used to find a value of u_0 at 1 m above the bottom. As the variation of u_0 with depth is unknown, we may only make a rough guess at the order of magnitude of the residual current at that depth. A velocity of 0.5 cm/s was found for northerly winds at a minimum height of 10 m above the bottom. Here we assume the same value, 0.5 cm/s as an upper value of u_0 at 1 m above the bottom. The drag coefficient k is taken 10^{-2} for an upper estimate of the bottom stress.

So for the absolute value of τ_{bo} we find:

$$\tau_{bo} \leqslant 0.013 \text{ kg/m}.\text{s}^2$$

A lower estimate is $\tau_{bo} = 0$ (stagnant bottom conditions).

In the equilibrium situation the combined effect of the surface stress and the bottom stress must be compensated by the pressure gradient, caused by the slope of the sea surface (if horizontal density gradients are neglected). This gives

$$\rho_w g\, d \frac{\mathrm{d}\, h}{\mathrm{d}\, x} = \tau_{xs} - \tau_{xb}$$

where the coordinate x is taken along the axis of the estuary (positive towards the mouth) and the index x means the components in the x direction.

A representative value for the depth of the inner and central ria may be obtained by taking the 'harmonic average'

$$\bar{d} = \frac{1}{\dfrac{1}{L}\displaystyle\int_0^L \frac{x}{d}\,\mathrm{d}x}$$

(see for instance GROEN, 1969). Here we use the average values of the depth for the different segments into which the ria has been divided (see Table XVII) and calculate d as the harmonic mean of the mean depths of these sectional averages. For ria segment B to F we find:

$$\bar{d} = 13.5 \text{ m}$$

For τ_s and τ_{bo} we take the upper and lower estimates given above, the values of τ_{bo} being negative, as the residual current near the bottom in the stationary situation is opposed to the wind direction. We find

$$0.4 \times 10^{-6} < \frac{\mathrm{d}h}{\mathrm{d}x} < 1.6 \times 10^{-6}$$

giving a total effect on the sea level over the length of the inner and central ria between 0.8 and 3 cm.

10.3 The influence of the density differences

The density differences along the ria, caused by differences of salinity and temperature, give rise to horizontal pressure gradients that cause a circulation of the water. The pressure at a certain depth below the sea surface is equal to the atmospheric pressure plus the weight per unit area of the overlying water column, which is found by vertical integration of the specific weight, In oceanography it is usual to express the vertical variation of the pressure not as a function of depth, but as a function of geopotential (which is given by the depth multiplied by the local acceleration of gravity). So the difference in geopotential of two levels lying 1 m apart is about 0.98 the unit for geopotential; this is called a dynamic metre.

Taking the average value of the observations of σ_t at the 3 positions A, B and C in the outer, central and inner ria, respectively (see the positions in fig. 1.2 and the curves for σ_t in fig. 6.4) we find the following values for the water pressure at different 'dynamic depths' below the local sea level:

TABLE XVI. *Pressure distribution in the Ria de Arosa*

geopotential		pos. A	pos. B	pos. C
sea level - 0 dyn. metre	pressure	0 decibar	0 decibar	0 decibar
- 5		5.132	5.131	5.113
- 10		10.265	10.263	10.246
- 15		15.399	15.396	15.380
- 20		20.534	20.530	20.515
- 30		30.805	30.799	
- 40		41.077	41.069	
- 50		51.349	51.339	

From these values the horizontal pressure gradient might be calculated if the slope of the sea surface between these positions were known. If the sea surface at the different points were equal, the pressure gradient would cause an inward flow of water at al depths, whereas, if the sea level at B were 1 cm higher than at A and the level at C were 1.5 cm higher than at B, there would be an outward flow at all depths, provided there were no wind stress and taking into account that the lateral constriction by the coasts opposes the development of geostrophic currents in that direction. An equilibrium between inflow and outflow exists in the case of no surface stress if the slope of the sea surface is somewhere between these two cases.

Thus, the slope of the sea surface estimated for the purely wind-driven circulation and for the purely density-driven circulation are of the same order of magnitude, be it of opposite sign.

10.4 General remarks on the theory of estuarine mixing and flushing

In the study of estuarine hydrography we may encompass the study of the dynamics and the flushing of a large variety of semi-enclosed coastal areas such as estuaries in the more limited sense of the word, fiords and lagoons (PRITCHARD, 1962). According to the salinity being higher, lower, or (about) equal to that of the adjoining sea area these coastal embayments are sometimes classified as negative or inverse, positive or neutral estuaries. The physical causes for movement and mixing of water may be different. Tide, wind, and pressure gradients (by river run-off or by thermal differences) may play a part.

The Ría de Arosa apparently is a positive estuary. Notwithstanding the low river discharge there is an excess of fresh water inflow over evaporation and the salinity increases from the head to the mouth of the ria. Only in its secondary bays different conditions may prevail, as is shown in section 6.17.

The salt balance at a given point in the estuary may be expressed by the relation

$$\frac{\partial S}{\partial t} = -u\frac{\partial S}{\partial x} - v\frac{\partial S}{\partial y} - w\frac{\partial S}{\partial z} + \frac{\partial}{\partial x}\left(K_x\frac{\partial S}{\partial x}\right) + $$
$$+ \frac{\partial}{\partial y}\left(K_y\frac{\partial S}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial S}{\partial z}\right)$$

Here x, y and z are the longitudinal, lateral and vertical coordinates of an orthogonal system, pointing downstream, to the right, northwest and upwards, respectively. The current components in the x, y and z directions are u, v and w, and K_x, K_y, K_z are the coefficients of eddy diffusion in the three directions. If the mean values averaged over one or more tidal periods are considered, u, v and w are the components of the residual current and the coefficients of eddy diffusion contain the net effect of the tidal movements on the dispersal of dissolved substances.

Usually we may neglect some of the advective and diffusive terms in this equation, depending on which of the processes of circulation and mixing is predominant. Another simplification is to suppose a stationary salinity distribution $\left(\frac{\partial S}{\partial t} = 0\right)$. Such more or less drastic simplifications give the possibility to estimate the diffusion coefficients or the transport from salinity data and river run-off.

We may test the validity of the steady-state assumption for the present case in the following way. For different compartments or sections of the ria such as those of fig. 10.3 the mean salinity was calculated from the observations in these sections (taking into account the depth distribution of the observations). For particulars of the mean salinities, etc. see fig. 10.3 and Table XVIII. Using $S = 35.66\ ^0/_{00}$ as the salinity of the inflowing ocean water, the contents of freshwater of the ria may be found from these mean salinities and the volume of each section. The inner and central ria together are estimated to contain about 18 × 10^6 m^3 of freshwater.

The discharge of the Río Ulla was estimated to be 11.5 m^3/s or about 1.0 × 10^6 m^3/day (section 3.4). Evaporation from the inner and central ria was estimated to be about 47 × $10^3 m^3$/day, which is negligible compared with the run-off. Precipitation is not estimated, but will also be relatively unimportant. We therefore may conclude that the average residence time of river water in the inner and central ria is about 18 days. For the whole ria, containing about 28 × 10^6 m^3 of freshwater and a net river input of about 1.2 × 10^6 m^3/day (discharge of rivers Ulla and Umia), the mean residence time is about 23 days.

Within a time interval of 20 or 30 days the environmental conditions may change significantly. The average river discharge in June is estimated to be 1.8 times that of July, and the coastal upwelling may not yet have started in June, while also the predominance of the northerly winds is less in this month. Anyhow, a fully stationary situation is not very likely. However, no distinct longterm trend of the salinity with time could be established in our observations. As is shown in fig. 10.2, where the results of the regular observations at Punta Preguntoiro are shown, we see large variations, and a remarkable minimum in the middle of July, but on a whole the values fluctuate around the same average in the beginning of July as well as in the middle of August. Here we shall make the assumption of a steady state, which gives the possibility to estimate the circulation rate. It is thought that, as a first approximation, this will give fairly satisfactory results.

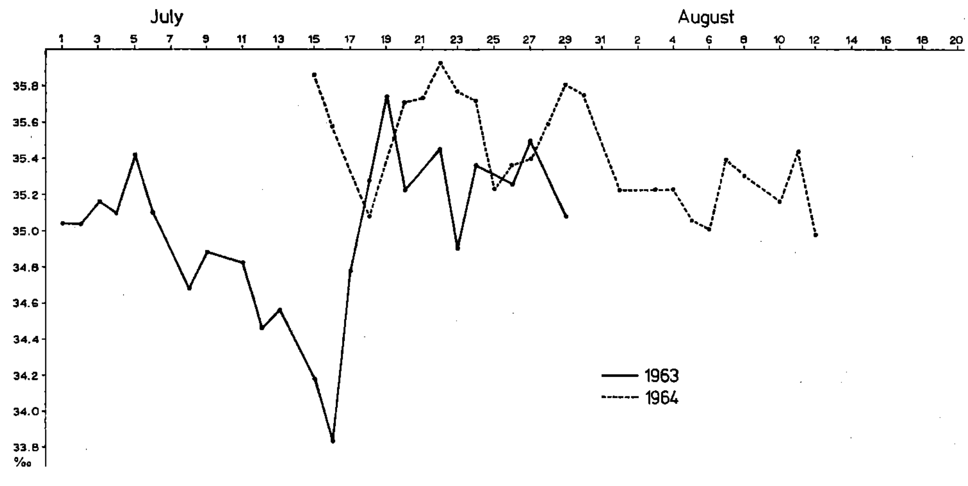


Fig. 10.2 Variation of salinity at Punta Preguntoiro.

10.5 Mixing and flushing of the Rio Ulla

The Río Ulla estuary is moderately stratified. Strong tidal currents give rise to turbulent mixing in a vertical as well as in a horizontal direction. No sufficient data are available to determine vertical and horizontal mixing and differences in the vertical up- or downstream advection. We can, however, investigate the flushing of the river by means of a one-dimensional model, in which the mean advective transport of salt through a cross-section is balanced by longitudinal turbulent transport. This turbulent transport in this case includes the longitudinal transport by the mechanism of vertical and lateral current shear. BOWDEN (1963) has shown that even in a vertically well-mixed estuary such as the Mersey estuary shear effects play an important part, and this can be expected to be even more so in a moderately stratified estuary such as the mouth of the Ulla before entering the ria.

The use of a one-dimensional model for the salinity transport implies that the general equation of section 11.4 is reduced to

$$\frac{\partial S}{\partial t} = -u\frac{\partial S}{\partial x} + \frac{\partial}{\partial x}K_x\frac{\partial S}{\partial x}$$

where S is now the mean salinity over a cross-section and u is the mean velocity of the current, while K_x is the apparent longitudinal eddy diffusivity.

If the average of the salinity over a tidal period is considered K_x will contain the net effect of the tidal movement on the dispersal of salinity, as mentioned before. If there is no long term variation of the salinity we may write $\frac{\partial S}{\partial t} = 0$. The cross-sectional mean of the residual current u is the mean river discharge R divided by D, the cross-sectional area of the river. So the equation is further simplified and becomes

$$0 = -\frac{R}{D}\frac{\mathrm{d}S}{\mathrm{d}x} + \frac{\mathrm{d}}{\mathrm{d}x}K_x\frac{\mathrm{d}S}{\mathrm{d}x}$$

Instead of this relation we may also take the integrated form. As the upstream or downstream transport of salt is zero in the (longterm) stationary case (neglecting salt advection by the river water) this relation takes the form

$$-\frac{R}{D}S + K_x\frac{\mathrm{d}S}{\mathrm{d}x} = 0 \text{ or } RS = DK_x\frac{\mathrm{d}S}{\mathrm{d}x}$$

if the contribution by the variation of the speed of the residual current along the estuarine part of the river $\left(\text{term } S\frac{\mathrm{d}}{\mathrm{d}x}\frac{R}{D}\right)$ is small enough to be neglected.

Comparison of this term with the advection term indeed shows a difference of about two orders of magnitude. This means that the assumption is justified.

For the Ulla river $R = 11.5\ \mathrm{m}^3/\mathrm{s}$ was used as a representative value of the summer discharge. In the cross-section half way between Catoira and Bamio the following values may be used for calculating the diffusivity:

$$S = 19\ ^0/_{00}, \frac{\mathrm{d}S}{\mathrm{d}x} = 6 \times 10^{-3}\ ^0/_{00} \text{ per metre}, D = 450\,\mathrm{m}^2$$

This gives $K_x = 80\ \mathrm{m}^2/\mathrm{s}$. Of course, this is a rough estimate. For the salinity distribution, as presented in fig. 6.11 (which was used for estimating the values of S and $\frac{\mathrm{d}S}{\mathrm{d}x}$), is based on only a small number of observations and observations over a complete tidal period are not available.

10.6 Circulation in the inner and central Ría de Arosa

In drawing up a quantitative model of the circulation of the ria we are mainly interested in large-scale phenomena, and in average values over a tidal cycle or more. Therefore we shall use a model based on a subdivision of the ria into a number of large segments. This model is an extension of that originally introduced by STOMMEL (1953) and used, amongst others, by DORRESTEIN AND OTTO (1960). Each of these segments covers about 2.5 km length of the main channel, a distance that is of the same order of magnitude as the tidal excursion.

The subdivision is shown in fig. 10.3. Boundaries between the segments are indicated by Roman numerals, the segments are indicated by capitals. The boundaries have been drawn in such a way that as much as possible the segments form topographically distinct units.

Since the circulation of the upper and deeper layers was shown to be quite different we shall use a model in which each segment has been divided into an upper and a lower layer. To that end we have to know the mean depth of the boundary-surface separating outflow and inflow. No direct observations in the form of current measurements along the vertical are available. So we have to use other evidence for deciding upon the depth of this boundary-surface.

In section 6.6 we have dealt with the vertical temperature structure. Apart from a temporary thermocline sometimes observed at very small depths (less than 4 m) another thermocline was observed as a regular feature, usually at a depth of about 10 m. We may assume that this thermocline coincides (or is very close to) the boundary we are looking for. This assumption is based on the following grounds.

1. Above the thermocline the vertical turbulent exchange of heat is apparently high,

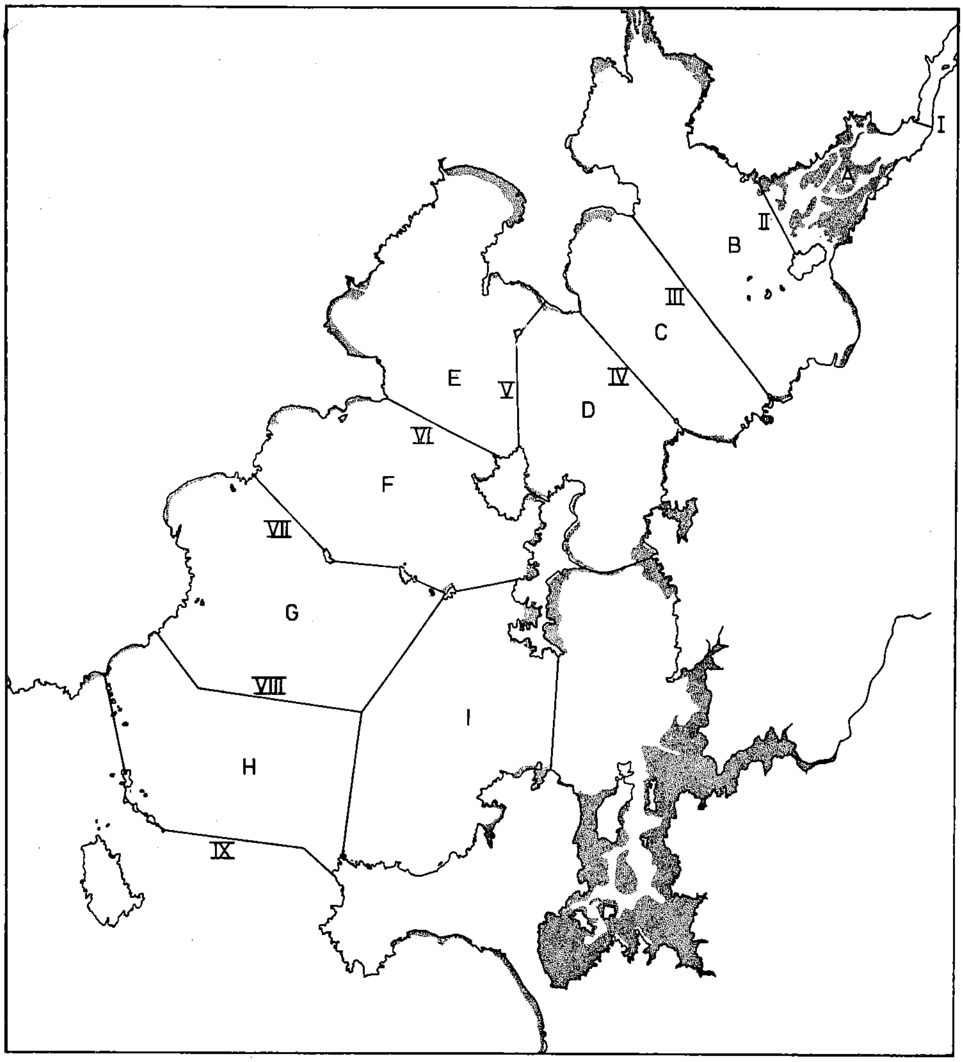


Fig. 10.3 Subdivision of Ria de Arosa for circulation model.

and the vertical eddy viscosity can be taken to be high as well. The current variations in the upper layer will therefore be rather small.

2. In the thermocline the density gradient is strongest, causing maximum static stability and minimum vertical exchange, so that strong current shear may be expected to exist here.

3. Below the thermocline the vertical variation of the temperature is small. This points to a common origin of the waters at these depths (oceanic inflow).

The depth distribution of the thermocline has been dealt with in section 6.6. It was mentioned there that sometimes a thermocline or an inflection of the temperature-depth curve occurred at more than one depth on the same registration. To determine the inflection point most likely to coincide with the boundary between upper and lower layer the following rule was used: if an inflection point occurred at less than 4 m depth together with another one at greater depth, the last one was assumed to show the level of the boundary surface, as the upper one can be ascribed to heating of the surface layers. In other cases in which more than one thermocline or inflection point was observed, the shallowest one was considered to give the right depth. In this way the mean depth of the boundary surface was estimated for the different segments into which the ria has been divided.

Now the mean salinity for the upper and deep layers of each segment could be determined from all observed salinity values. Only the regular observations at Punta Preguntoiro have been left out of consideration as these would give too strong a bias towards this special locality.

In fig. 10.4 the variation of the mean salinity along the main channel of the ria (segments A to H) is shown for both layers.

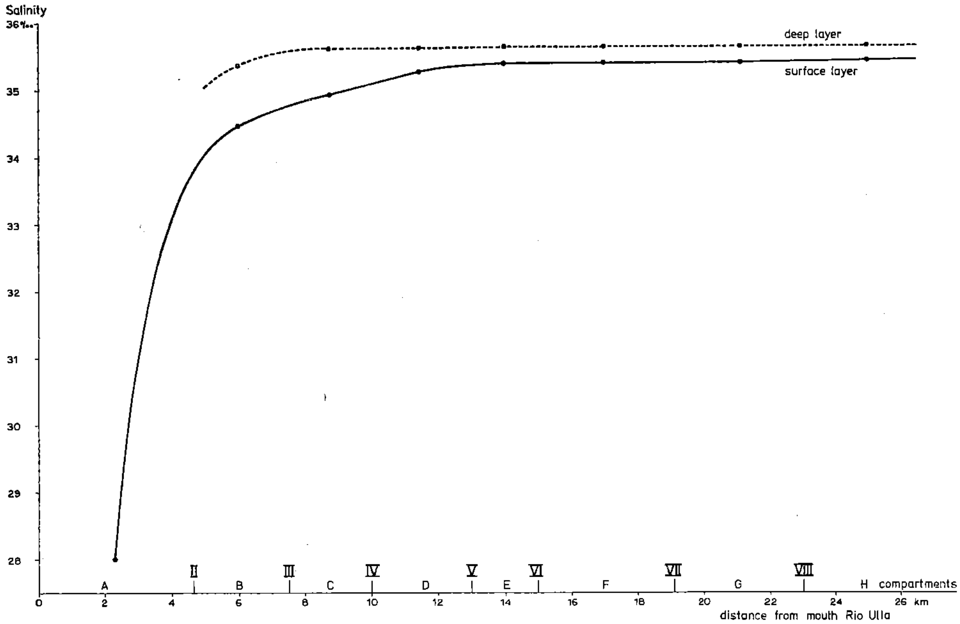


Fig. 10.4 Variation of mean salinity along the axis of the ria.

It can be seen that starting from segment C the mean salinity of the deep layer does not change significantly, and from segment E on the salinity of the surface layer is constant too. This indicates that no important exchange of water between both layers takes place further seaward of segment E and that vertical mixing, which lowers the salinity of the deep layer, is only slight, except in segment B.

These mean salinities are averages over values that may differ according to the location within the segments and because of tidal or wind-induced variations. In the mean values considered here we assume that tidal effects, wind effects and local variations are cancelled out because of the rather even distribution of the observations and the relatively small magnitude of the variations with wind and tide, discussed in section 6.12. Therefore we consider the equations for the transport of salt to apply to the average transport over a tidal cycle, as has been done in the previous section. Again this transport is supposed to be zero.

Because of the differences in the salinity and the current of the upper and lower layer we cannot integrate the equation given in section 10.4 for the salt balance at one point over a full vertical boundary between two sections, but we have to integrate over the upper part and the lower part of this boundary.

We now may write

$$R'S' + R''S'' - K'_x D' \left(\frac{\partial S}{\partial x} \right)' - K''_x D'' \left(\frac{\partial S}{\partial x} \right)'' = 0$$

where the symbols with one or two accents indicate the values of the upper or lower layer respectively, and the symbols have the same meaning as in the preceding section (see also fig. 10.5). Only R does not stand for the river discharge but the net residual water transport in either the upper or the deep layer. The use of a two layer model precludes largely the contribution of shear effects to the horizontal eddy diffusivity.

In deriving this expression it was also assumed that the variation of the residual current in the x direction is small, an assumption that appears to be justified afterwards, when considering the outcome of the computations with this model.

Furthermore, neglecting evaporation and precipitation, which is allowed in this case, the continuity equation for water indicates that the net transport of water through the whole boundary surface between two segments (upper and deep layer together) is equal to the river discharge.

$$R' + R'' = 11.5 \text{ m}^3/\text{s}$$

Except at the boundary between segment A and B where $R'' = 0$ (because segment A has no deep layer), we have to know the coefficients of horizontal diffusion K'_x and K''_x in order to solve the equations for the transport of water and of salt.

The exchange of salt between the deep and the upper layer of a segment may be

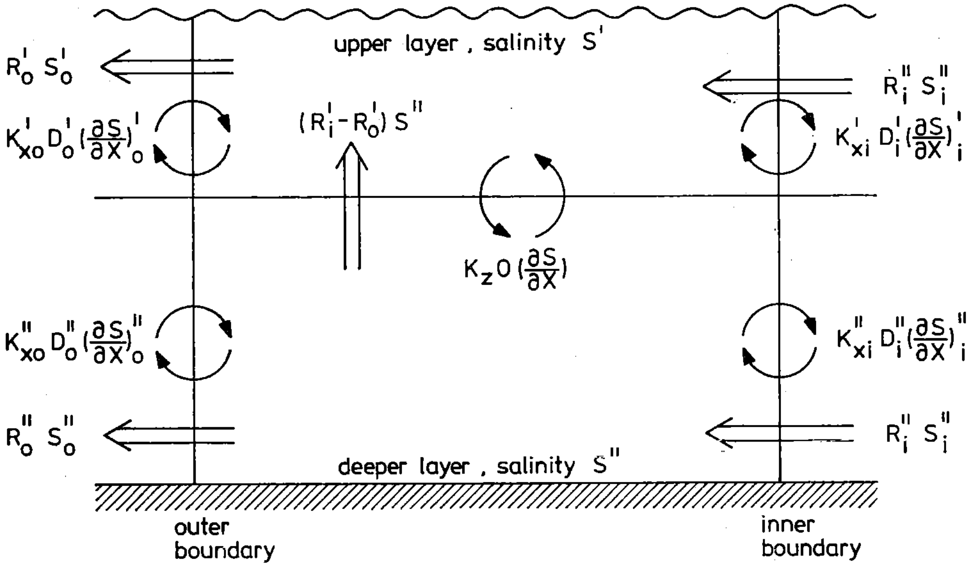


Fig. 10.5 Model of salinity exchange between different compartments of the ria.

found from the relation that considers the inflow and outflow of salt in the deep layer

$$R_0''S_0'' - R_i''S_i'' - K_{x0}''D_0''\left(\frac{\partial S}{\partial x}\right)_0'' + K_{xi}''D_i\left(\frac{\partial S}{\partial x}\right)_i'' = (R_0'' - R_i'')S'' - K_z O\left(\frac{\partial S}{\partial z}\right)$$

where the suffixes o and i indicate the values at the outward and inward boundaries of the segment, and the other symbols refer to the values of the segment itself (see fig. 10.5). O is the surface area of the boundary between both layers.

In table XVII the values of the salinity are given together with the relevant dimensional data of the segments and their boundaries. The depth of the boundary surface between upper and deep layer as estimated in the way indicated previously is given too. The values of the horizontal salinity gradients are approximated from the differences between the salinities at the successive segments

$$\left(\frac{\partial S}{\partial x}\right)'_{N-1,N} = \frac{S'_N - S'_{N-1}}{x_N - x_{N-1}}$$

where x_N is the distance along the main channel of the ria of the middle of section N from a certain reference point (here the head of the estuary).

TABLE XVII.

x	segment	segment	D'	D''	$\bar{d}$	d'	O_S	S'	S''	R'	R''	R_v	$\boldsymbol{K_x}'$	$\boldsymbol{K_x}''$	$\boldsymbol{K_v}$	$\bar{u}'$	$\bar{u}''$	w
10^3 m	boundary		10^3 m^2	10^3 m^2	m	m	10^6 m^2	‰	‰	m^3/s	m^3/s	m^3/s	m^2/s	m^2/s	10^{-6} m^2/s	cm/s	cm/s	10^{-3} cm/s
0	I		0.45	—				19.00	—	11.5	—		80	—		2.5	—	
2.32		A			3.3	—	7.7	28.00	—			—						—
4.65	II		10.0	—				33.80	—	11.5	—		22,5	—		0.1	—	
6.07		B			6.1	4	28.2	34.47	35.37			398.5			12			1.4
7.50	III		28.5	41.0				34.75	35.60	410	— 389.5		10	5		1.4	—1.0	
8.74		C			13.0	6	15.5	34.89	35.62			323			12			2.1
9.98	IV		32.0	46.0				35.10	35.63	733	— 721.5		10	5		2.3	—1.6	
11.46		D			16.8	10	18.5	35.27	35.65			612			10			3.3
12.95	V		34.2	50.4				35.37	35.66	1345	—1333.5		10	5		3.9	—2.6	
13.95		E			20.1	8	23.2	35.43	35.66			436			0			1.9
14.95	VI		29.5	84.7				35.43	35.66	1781	—1769.5		10	—		6.0	—2.1	
17.03		F			27.3	11	23.8	35.43	35.66			0			0			0
19.10	VII		68.0	77.6				35.43	35.66	1781	—1769.5		10	—		2.6	—2.3	

TABLE XVIII.

x 10^3 m	segment boundary	segment	t' °C	t'' °C	Q_H 10^8Watt	Q_V 10^8 Watt	$\triangle Q_1$	$\triangle Q_2$	$\triangle Q_{tot}$	$\triangle t$ °C/day
0	I		19.2		10					
2.32		A	18.4			—	— 3	+15	+12	+0.9
4.65	II		16.8		13					
6.07		B	16.6	15.2		245	—26	+56	+30	+0.5
7.50	III		16.5		284					
8.74		C	16.3	13.5		179	—24	+31	+ 7	+0.1
9.98	IV		15.9		487					
11.46		D	15.3	13.0		329	—59	+37	—22	—0.2
12.95	V		15.7		875					
13.95		E	16.0	13.0		236	—75	+46	—29	—0.3
14.95	VI		16.0		1186					
17.03		F	16.0	13.0		0	—21	+48	+27	+0.2
19.10	VII		16.3		1207					
									+25	+0.06

The value of $\frac{\partial S}{\partial z}$ is approximated from the difference between the salinity of the upper and that of the deep layer

$$\frac{\partial S}{\partial z} = \frac{S' - S''}{\frac{1}{2}d}$$

where d is the mean depth of the segment. This means that it is approximated by a segment with retangular profile. For the surface area of the boundary layer O we therefore have to take the surface area of the segment O_s.

As stated earlier we have to know the coefficients of horizontal turbulent diffusion for the upper and lower layers, K'_x and K''_x for the different segments (except for the one-layered segment A). The eddy diffusion depends on the scale l of the transport process that is considered, which in our case means the length scale of the segments into which the ria has been divided (about $2\frac{1}{2}$ km). In the open sea, according to OKUBO (1971) we may expect a relation for the relation between scale and eddy diffusivity that is given by

$$K = 0.0103\, l^{1.15}$$
$$(K \text{ in cm}^2\text{/s and } l \text{ in cm})$$

This relation would give a value of K of about 1.8 m^2/s with $l \doteq 2.5 \times 10^5$ cm. In tidal estuaries, however, much higher values of K are found because of the effect of tidal mixing and shear effects. For the vertically well-mixed Ems estuary on the German-Dutch border DORRESTEIN AND OTTO (1970) found values for K of the order of 100 m^2/s, using a calculation method based on a subdivision in segments of a length scale of 5 to 10 km. Values of K between 50 and 350 m^2/s have been found in comparable situations (BOWDEN, 1963). The apparent eddy diffusion found for the Rio Ulla (section 10.5) falls in the same range of values (K = 80 m^2/s). In all these situations, however, the vertical and lateral shear plays an important part in the mixing process. In a two-layered circulation model as used here the effect on K of the vertical shear is quite different from that of the other estuarine models mentioned above, and smaller values may be expected for the K values of each layer.

Published data on the horizontal eddy diffusivity of surface layers in deep embayments or estuaries are scarce. GADE (1968) estimated the horizontal diffusion coefficient of the main part of the Oslo Fjord to be of the order of 100 m^2/s. Solving the salt balance equations for the one-layered segment A of the ria a value for K_x of 22.5 m^2/s was found. In the circulation model combinations of different values for K'_x, and K''_x were tried (K'_x = 5, 10, 25 and 50 m^2/s and K''_x = 0 and 5 m^2/s). For K'_x = 50 m^2/s inconsistent results were obtained (negative values of the vertical eddy diffusivity in some of the sections). The other values of K'_x and K''_x all give a more or less acceptable circulation model (in that case the R values found are roughly in agreement with observed currents). As a criterion for the final choice between the different possibilities the results obtained in the heat budget were used (section 10.7). Those K values were chosen which give a heat balance that is as close as possible to equilibrium. It was found that the best results were obtained with K'_x = 10 m^2/s and K''_x = 5 m^2/s for all segments except E, where the only possible value usable for K''_x is zero.

The results of the calculations are given in table XVII where the transport of water between the different segments and the upward transport within a segment are given. As may be seen from this table the mean current velocities of the surface layer are about a third to a quarter of the velocity of the residual current for northerly winds as deduced from our current observations (see section 6.10). This discrepancy, however, is not so serious if we realize that in the first place the observed values have a large standard deviation, that furthermore the mean current is the average over a layer of about 10 m thickness while the observed currents pertain to a near-surface layer of about 1 m depth, and that also the balance for salt and water applies to the average summer situation, during which south-westerly winds also occur regularly.

It is interesting to note that the vertical water transport is at a maximum in segment D, which is the area where, according to temperature, nutrient and oxygen data the upwelling is most important.

The renewal of the deep water layer, according to these figures, is a rather rapid process, it takes about one week to have a complete renewal of the waters in the deep layers of the inner and central parts.

For the Ría de Vigo MARGALEF AND ANDREU (1958) estimated the vertical movement of the water and the vertical eddy diffusivity from the time variations of heat content below a certain level. In their calculations the horizontal advection of heat is neglected. The estimated eddy diffusivities are much higher than those found here. They found ascending motions of the deep water with velocities of the same order of magnitude. But because of the somewhat elementary calculations these values, in our opinion, cannot be used for a quantitative comparison.

As regards the circulation in the remaining parts of the ria there are not sufficient data for a quantative evaluation of the complex system that may be expected from the information discussed in chapter 6. The sectionally mean salinity over a segment is virtually constant for both the upper and the deep layers of the outer ria. Exchange between both layers therefore is considered to be negligibly small compared with horizontal exchange. Indications have been found of a surface inflow along the eastern side of the main channel. It is not quite clear what happens to this inflow. Exchange may occur with the outflowing water at the western side, or with water from the southeastern part.

Also the exchange in the southeastern part where the outflow of the Rio Umia is important is difficult to evaluate. The exchange over a shallow sill probably is an intermittent process and should be observed regularly for some time for a quantitative assessment to be possible. A further complicating factor is the influence of the shallow bay of El Grove. Because of all these uncertainties we will abstain from an attempt to estimate the water exchange in this region quantitatively.

10.7 Heat budget of the central and inner ria

The heat budget of the sea consists of two parts: the amount of heat transported into and out of the area considered by advective and diffusive processes ($\triangle Q_I$), and the heat exchange through the sea surface by radiation and exchange with the atmosphere ($\triangle Q_{II}$). Contributions by heat flow through the sea bottom, by dissipation of kinetic energy of waves and currents and from chemical processes may be neglected in the majority of the cases.

The first part of the heat balance mentioned above can be calculated for the inner and central part of the Ría de Arosa using the circulation model given in the previous paragraph. The second part, the exchange through the sea surface has to be estimated from different meteorological and oceanographical parameters.

The terms of the heat balance that make up the exchange through the sea surface are:

1. Q_s: total incoming radiation
2. Q_r: reflected radiation
3. Q_b: net radiation loss from the sea surface being the difference between Q_{bs},

the long-wave radiation emitted by the sea surface and Q_{ba} the back radiation from the atmosphere

4. Q_e: heat lost by evaporation (or gained by condensation)
5. Q_h: convection of sensible heat to and from the atmosphere.

In practical studies of the heat balance the estimates of the different terms have to be made from standard meteorological measurements. For a more fundamental approach measurements at several levels that should be available usually are lacking. However, empirical relations based on certain reasonable assumptions on the nature of the transfer processes can be used.

The incoming radiation may be calculated using the tabulated values for solar radiation at the outside of the earth's atmosphere such as found in the Smithsonian Meteorological Tables (LIST, 1951). According to these figures the incoming radiation outside the atmosphere for latitude 42° 30′ N and the month July is on the average 910 cal/cm^2 day (440 Watt/m^2).

This radiation is partly absorbed or scattered by the atmosphere. The atmospheric turbidity, the presence of clouds and the sun's altitude together determine the incoming radiation at the sea surface Q_s. Empirical relations of the form

$$Q_s = I_0 \{a + b(1 - C)\}$$

are often used to calcultate the incoming radiation from the value outside the atmosphere I_0 for a certain cloudiness C. The coefficients a and b contain the effects of average local atmospheric conditions and of the latitude and therefore they may vary from place to place and from one month to the other. For the situation near the Ría de Arosa we may use the values given by BARATA (1955) for Lisbon. According to this author the coefficients for July should be

$$a = 0.19$$
$$b = 0.71$$

With the mean cloudiness observed during the summer campaigns ($C = 0.44$) this gives

$$Q_s = 540 \text{ cal/cm}^2 \text{ day (260 Watt/m}^2\text{)}$$

The error in Q_s is difficult to estimate but judging by other studies it is thought to be about 10%. Part of this radiation is reflected by the sea surface while furthermore some of the radiation penetrating the sea is scattered backwards. The percentage of reflected radiation depends on the distribution of the radiation between direct and diffuse radiation from sky and clouds and the roughness of the sea surface, while also the variation of the solar altitude during the day is of importance.

JERLOV (1968) has summarized recent literature on this subject. According to this review the reflectance for direct radiation is about 2 % around midday and gradually increases if one considers earlier or later hours of the day. However, for a rough sea the reflection will not be more than 16 % (at solar altitude 10°). The net reflection for diffuse radiation varies between 4.3 and 6.6 % for different conditions. As the diurnal variation of incoming radiation has a maximum at high solar altitudes, we use an effective reflection coefficient for direct radiation of 6 % (in agreement with the estimate given by SELLERS, 1965) while for the reflection coefficient of diffuse sky radiation we equally use a value of 6% . An over- or underestimate of one or two percent, however, has no great effect on the final outcome of the heat balance because the reflection, and especially that of the diffuse radiation, is only of secondary importance. So for the reflected radiation we find

$$Q_r = 32 \text{ cal/cm}^2 \text{ day } (16 \text{ Watt/m}^2)$$

The value thus found for $Q_s - Q_r = 508$ cal/cm^2 day is in good agreement with the figure reported for the Ría de Vigo by VIVES AND FRAGA (1961[A]).

The long wave radiation emitted by the sea surface is only slightly less than that of a black body of the same temperature. A value of 0.97 for the emissivity of sea water for a representative temperature of 290 °K (17 °C) gives a longwave heat emission $Q_{bs} = 800$ cal/cm^2 day (390 Watt/m^2). Systematic diurnal variations of the water temperature at Punta Preguntoiro are less than $\pm$ 1.5° C, and we may fairly assume that for the major part of the ria this will be smaller. This means that the average value of Q_{bs} for a whole day will differ no more than 1 % from the average value at a particular time of the day.

Back radiation from the atmosphere mainly comes from the lowest layers. For this reason this back radiation may be estimated from near-surface meteorological observations. Because of the emission of infrared radiation by water vapour the moisture of the air is an important parameter in formulas used for estimating back radiation. Back radiation from the atmosphere at a clear sky can be estimated with the often used Brunt equation

$$Q_{ba} = \varepsilon \times 11.7 \times 10^{-8} T^4 (\boldsymbol{a} + \boldsymbol{b}\sqrt{e})$$

where ε is the emissivity, and T and e are the temperature and vapour pressure at a certain measuring level in °K and millibars respectively. The coefficients $\boldsymbol{a}$ and $\boldsymbol{b}$ are empirically determined. They will vary somewhat with the level for which T and e are measured. In our situation the diurnal varitation of the mean vapour pressure is small (between 16 and 17.5 mb), so it can be expected that it is possible to estimate the back radiation with sufficient accuracy from meteorological data from 9 and 21 h local time, when the deviation from the diurnal mean is only small.

SELLERS (1965) gives as the median values of a great number of estimates for the Brunt coefficients

$$a = 0.605$$
$$b = 0.048$$

In the following we will use these values together with an emissivity 1. Because the vapour pressure is close to that of the water any error resulting from application of data from a level for which the coefficients have no validity will come from the difference in temperature. Considering the difference between the water temperature and the mean air temperature we estimate this error to be no more than 1 %.

The presence of clouds decreases the radiation loss by increasing the back radiation. Sellers gives the following relation for the back radiation under cloudy conditions:

$$Q_{ba} = Q_{ba\,(\mathrm{clear})}\,(1 + k_1 C^2)$$

where k_1 varies from 0.04 for Cirrus clouds to 0.25 for fog. For the Cumuli-form clouds most frequently observed we may take $k_1 = 0.2$. If we use the mean cloudiness of the daytime observations, that is 0.44, we would find

$$Q_{ba} = Q_{ba\,(\mathrm{clear})} \times 1.04$$

If during the night, when no observations were made, the sky would have been completely overcast or clear, this would have meant a multiplication factor of 1.08 or 1.02 respectively. So, using the daytime cloudiness data only we would make an error of at most 4 %.

The heat lost by evaporation can be estimated from the data given in section 3.5 and the latent heat of evaporation (587 cal/cm^3 at 17 °C water temperature). Although a very accurate estimate of the evaporation appeared to be beyond the possibilities of this study it was shown that in any case the evaporation was rather low. A value of 0.4 mm/day was found to be of the right order of magnitude.

The exchange of sensible heat by convection Q_h is estimated from the difference in temperature at the sea surface and at some higher level, using a transfer coefficient for heat, D_h

$$Q_h = \rho_{\mathrm{air}} c_p D_h (t_s - t_z)$$

where c_p is the specific heat of the air at constant pressure ($c_p = 0.24$ cal/g. °C).

Often it is assumed that the transport processes for heat and moisture in the near-surface layers are equal ($D_w = D_h$). In section 3.5 we assumed a value for D_w based on empirical evaporation studies. Although it is not certain whether the assumption $D_w = D_h$ completely holds for the layer thickness considered here, it is thought that the estimates thus obtained will not be much in error. This is supported by the observation of ANDERSON (1954) that the estimates based on this assumption in general give consistent results.

As regular meteorological observations were not made during the night we have to estimate the error that is made in Q_b and Q_h by using day-time observations only. For the evaporation this problem was already dealt with in section 3.5, where the error found to be small. For the two occasions on which observations were made over a period of 24 hours, also used in the discussion of the evaporation estimate, the following results were obtained:

July 11-12, 1963 (SW winds, Punta Preguntoiro). Units: cal/cm^2. day.

estimate from:	8 3-hourly observations	$Q_{ba\,(\mathrm{clear})} =$	694	$Q_h =$	-54
	mean of 8 3-hourly obs.		<u>696</u>		<u>-38</u>
	09 and 21 h observations		698		-41
	mean of 09 and 21 h obs.		698		-39

July 19-20, 1963 (N winds, La Toja). Units: cal/cm^2. day.

estimate from:	8 3-hourly observations	$Q_{ba\,(\mathrm{clear})} =$	731	$Q_h =$	-155
	mean of 8 3-hourly obs.		<u>710</u>		<u>-152</u>
	09 and 21 h observations		725		-200
	mean of 09 and 21 h obs.		722		-194

The formula used is intended to give estimates from diurnal means of the meteorological parameters. So the underlined values are supposed to be the 'best' estimates. It can be seen that the different estimates are very close together for Q_{ba}, while for Q_h the differences are greater and are about 40 % and 30 % of the 'best' estimates, respectively. Although these latter differences are important as such, they have a relatively small influence on the final result of the total heat input via the sea surface. Therefore, judging from these findings we feel that it is possible to obtain a reasonable estimate for the values of Q_{ba} and Q_h from the 09 and 21 h local time data. A rough estimate of the variation of the heat exchange over the ria, using the regional differences with Punta Preguntoiro mentioned in section 2.6, shows that the effect of those differences on the total heat exchange is relatively small, giving differences of less than 10 %. This is mainly because in the most important term in which regional atmospheric differences play a part, Q_{ba}, temperature differences as observed over the ria (about 2 °C) give rise to differences in Q_{ba} of only some 20 cal/cm^2 day.

Therefore these regional differences, being within the limits of the accuracy of the estimates, have been neglected. From the data available we estimated the following values:

Heat input through the sea surface:

Q_s = 540 cal/cm^2. day (260 Watt/m^2)
Q_{ba} = 705 cal/cm^2. day (340 Watt/m^2)
Q_h = 444 cal/cm^2. day (21 Watt/m^2)

Heat output through the sea surface:

Q_r = 32 cal/cm^2. day (16 Watt/m^2)
Q_{bs} = 800 cal/cm^2. day (390 Watt/m^2)
Q_e = 24 cal/cm^2. day (11 Watt/m^2)

So the input is about 620 Watt/m^2 and the output 420 Watt/m^2 giving a net heat transport through the sea surface (inward) of 200 Watt/m^2.

The advective and diffusive transport of heat (resulting in a net heat flow $\triangle Q_l$) will be calculated with the model presented in section 10.6. The mean temperature of the upper layer (not to be mistaken for the mean temperature of the surface, which is usually somewhat higher) and the mean temperature of the deep layer are given in figure 10.6.

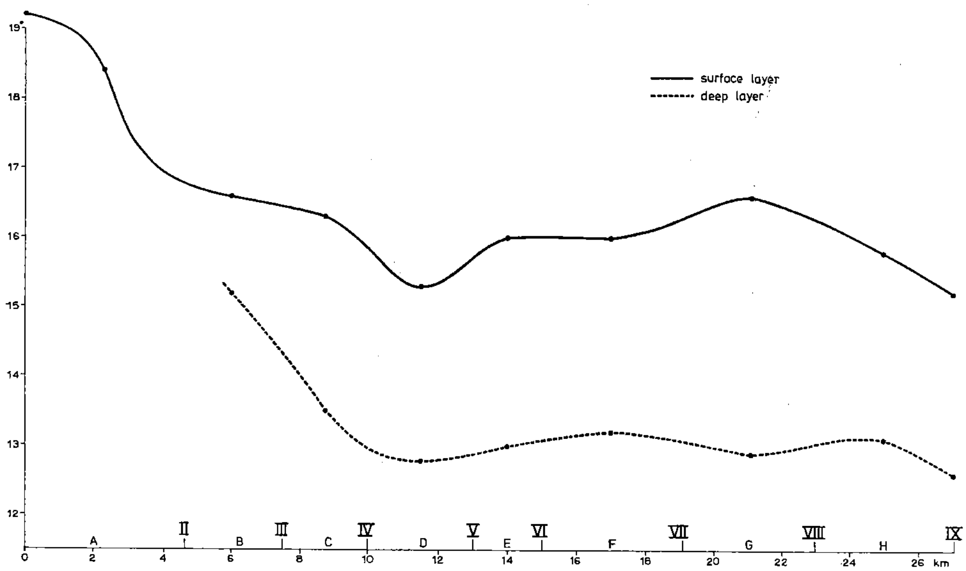


Fig. 10.6 Variation of mean temperature along the axis of the ria.

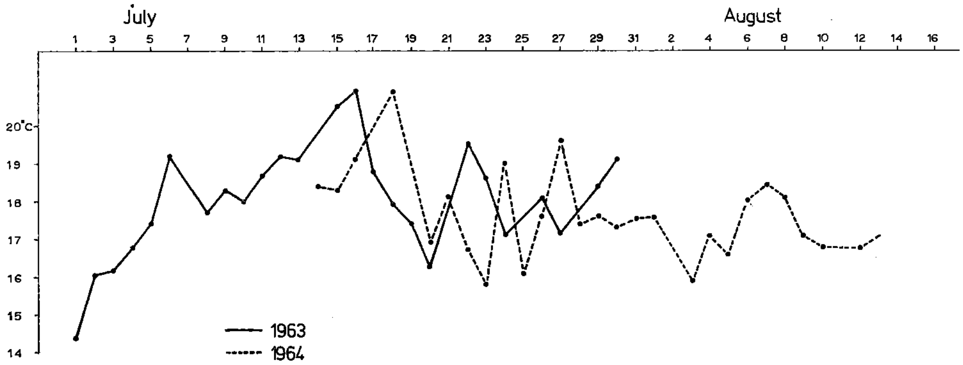


Fig. 10.7 Variation of temperature at Punta Preguntoiro.

In Table XVIII (p. 166) the results of the calculations are presented for the upper parts of the segments A to F of the inner ria. The specific heat of the water is taken to be 0.93 cal/cm^3. For each individual segment there are large residues, which would give a considerable warming up or cooling down of the surface layer. From the observations at Punta Preguntoiro we may conclude that temperature variations of the order of 0.2 °C/day may occur, apart from short-time variations (see fig. 10.7). Apparently the calculations of the heat exchange with the atmosphere ($\triangle Q_{II}$) or of the transport in the water ($\triangle Q_I$) have deficiencies. It is likely that the large residue for segment A is due to the presence of areas with very shallow water and tidal flats which may influence the heat belance in a way that is not accounted for in the foregoing calculations.

The overall heat balance of the whole area, however, is quite satisfactory. This gives confidence in the model as a way of calculating the transport of different substances or properties in and out of this area as a whole. The results obtained for the different segments should be considered to give more or less relative trends within the area of the central and inner ria rather than absolute values.

10.8 The interrelation between the dynamics of the circulation and the salinity balance

In a series of publications Rattray and Hansen have given an approach to the problem of the interrelation between the dynamics of the circulation and the salinity distribution in estuaries. In these studies the density and the salinity are considered to have a linear relationship, an assumption which does not need to be true if heat is exchanged with the atmosphere, and which does not hold in every respect for the Ría de Arosa in summer.

In a first study (RATTRAY AND HANSEN, 1962) these authors use a model in which among other things they neglect the vertical advective salt flux, which means that this approach cannot be applied to the inner and central parts of the ria, where upwelling takes place. However, it might be useful to apply this model to a study of the processes of circulation and mixing in the outer ria. In two other studies (HANSEN, 1967; and HANSEN AND RATTRAY, 1965) the theory includes vertical advection. Two cases are considered by these authors: one of a salinity which increases linearly with the distance along the estuary, and another of an exponentially increasing salinity; these cases are considered to be typical for the regime in the central and in the inner parts of an estuary.

Our interest is especially with the central regime. In their theory Hansen and Rattray have to make several assumptions. The coefficients for vertical eddy diffusivity and eddy viscosity are taken to be constant.

Finally the estuary is considered to be a channel with constant cross-section.

Although the assumptions made are somewhat removed from the situation as we find it in the Ría de Arosa, we shall investigate in how far this second model (linearly increasing salinity) agrees with our mesaurements in the inner ria, represented by segment C. The longitudinal salinity gradient in the segments B, C and D can be approximated with a linear increase of about 10^{-4} $^0/_{00}$ per metre, as may be seen from fig. 10.4. We shall follow the theory as given in the publication of HANSEN AND RATTRAY (1965).

The following values will be used:

mean depth (see Table XXI)	$d = 13$ m
mean width	$B = 5.10^3$ m
wind stress (see section 11.2)	$\tau_w = 0.05$ kg/m. s^2

As regards the relation between salinity and density Hansen and Rattray assume a linear relationship of the form

$$\frac{\rho_1}{\rho_2} = \frac{1 + \mathrm{k}S_1}{1 + \mathrm{k}S_2}$$

Here we consider the water of the ria as a mixture of North Atlantic Central water and the so-called ria surface water, for which we shall use the average values $S = 35.75$ $^0/_{00}$, $t = 12.4$ °C, $\sigma_t = 27.11$ and $S = 35.20$ $^0/_{00}$, $t = 17.4$ °C, $\sigma_t = 25.62$ respectively (see section 6.3). This gives for k a value of 3.10^{-3} $(^0/_{00})^{-1}$.

Furthermore we use for the rate of river discharge (see section 3.4) a value

$$R = 11.5 \text{ m}^3/\text{s},$$

for the horizontal diffusion coefficient at $x = 0$ (see section 10.6) we use the surface value

$$K_{h0} = 10 \text{ m}^2/\text{s}$$

The dimensionless vertical and longitudinal co-ordinates are

$$\eta = \frac{z}{d} = \frac{z}{13 \text{ m}}$$

$$\xi = \frac{Rx}{BdK_{h0}} = \frac{x}{56 \text{ m}} 10^{-3}$$

Rattray and Hansen furthermore use a gradient parameter ν, that is defined by $\nu = \dfrac{1}{S_0} \dfrac{\partial S}{\partial \xi}$ (here S_0 is the vertical mean salinity at $x = 0$), a dimensionless wind stress

$T = \dfrac{Bd^2\tau_w}{A_v \rho R}$, an estuarine Rayleigh number $R_a = \dfrac{gkS_0 d^3}{A_v K_{h0}}$

and a tidal-mixing parameter $M = \dfrac{K_v K_{h0} B^2}{R^2}$, where A_v and K_v are the vertical eddy viscosity and eddy diffusivity, respectively.

Using $S_0 = 35.25$ (the mean salinity at the upstream boundary of section C, see table XVII) and the other values given above, we find

$$\nu = 0.16$$

$$T = \frac{3.7 \text{ m}^2/\text{s}}{A_v}$$

$$R_a = \frac{23 \text{ m}^2/\text{s}}{A_v}$$

$$M = 1.9 \cdot 10^6 \text{ m}^2/\text{s} \times K_v$$

The value of A_v can be estimated from the estimated water transport in the 6 m thick upper layer of section C. Calculating from the data of table XVII the mean value of the transport Q' at the upstream and downstream boundary we find

$$Q' = 572 \text{ m}^3/\text{s}$$

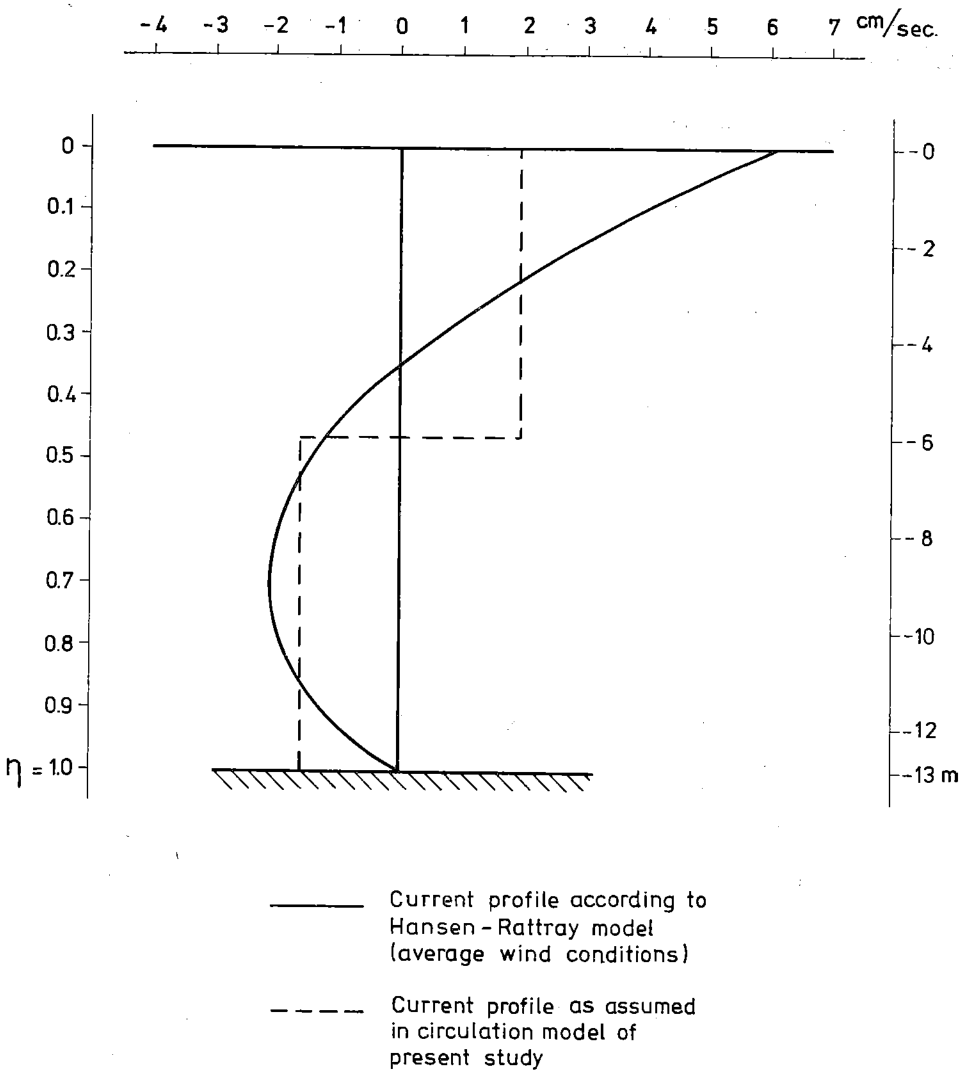


Fig. 10.8 Current profile according to Hansen-Rattray theory, mean conditions.

With a stream function $\varphi(\eta)$ as defined by Hansen and Rattray we can write for the water transport between the sea surface and a depth η'

$$Q(\eta') = R\{1 - \varphi(\eta')\}$$

Substituting in Hansen-Rattray's formula for the stream function the values $Q' = 572 \mathrm{m}^3/\mathrm{s}$ and $\eta' = 0.46$ ($z = 6$ m) we find with the parameters as given above

$$A_v = 2.95 \times 10^{-3} \ \mathrm{m}^2/\mathrm{s}$$

and this gives for $\boldsymbol{T}$ and $\boldsymbol{R}_a$

$$\boldsymbol{T} = 1.25 \cdot 10^3$$

$$\boldsymbol{R}_a = 7.8 \cdot 10^3$$

The velocity profile that follows from these parameters according to the theory is given in fig. 10.8. The surface velocities found are about 6 cm/s, that is of the right order of magnitude (see section 6.10). According to the theory the current reverses at a depth of 4.5 m, this is 1.5 m less than the estimated depth of the reversal.

The value of the tidal mixing parameter $\boldsymbol{M}$ can be estimated from the mean salinity of the upper layer of segment C, $S' = 34.89$. Integrating Hansen and Rattray's expression for the salinity as a function of ξ and η between $\eta = 0$ and $\eta = 0.46$ ($z = 6$ m) we find for a point in the centre of segment C, at $\xi = 0.022$,

$$S' = \frac{35.25}{0.46}\left(0.478 - 4.8 \ \frac{\nu}{\boldsymbol{M}}\right) = 34.89$$

We find $\boldsymbol{M} = 34$ and $\boldsymbol{K}_v = 18.10^{-6} \mathrm{m}^2/\mathrm{s}$, which differs 50% from the value estimated from the circulation model ($\boldsymbol{K}_v = 12.10^{-6} \ \mathrm{m}^2/\mathrm{s}$).

However, notwithstanding these discrepancies, the agreement between both models is considered to be acceptable.

We therefore believe that the theory can be used in the investigation of the consequences of deviations in the environmental conditions from the average situation. The nature and order of magnitude of the resulting changes in the circulation pattern can be evaluated.

Such a change could be the occurrence of wind conditions significantly different from the mean, such as either fresh northerly or southerly winds or calm conditions.

Assuming that the different parameters as river discharge, horizontal and vertical eddy diffusivity remain the same, and furthermore assuming that the gradient parameter ν does not change significantly (which means that we assume the departure from the average situation to be of not sufficiently long duration to alter the salinity distribution appreciably) we can investigate the effect of a change in $\boldsymbol{T}$ on the system.

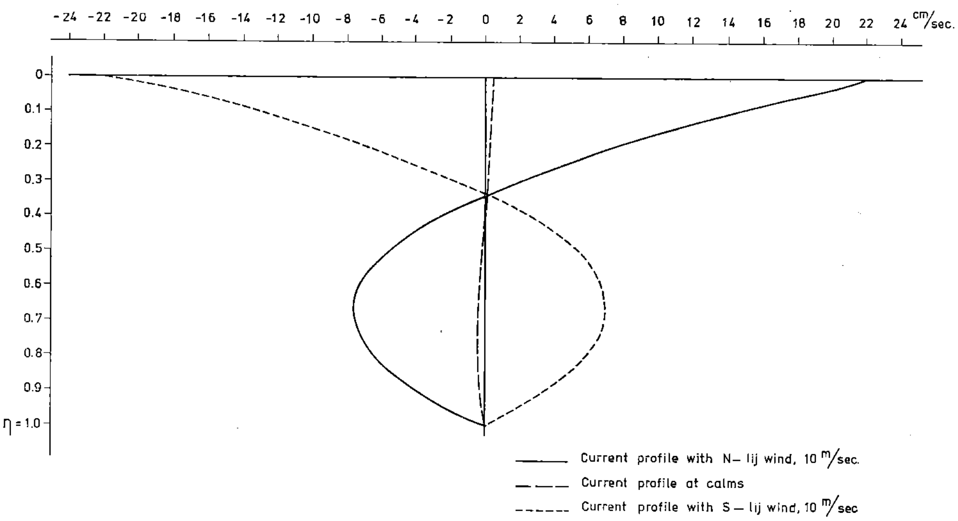


Fig. 10.9 Current profile according to Hansen-Rattray theory, different wind conditions.

For wind speeds of 10 m/s we find a wind stress of about 0.2 kg/m s^2. Thus, introducing

$T = 5.0 . 10^3$ (northerly wind)
$T = 5.0 . 10^3$ (southerly wind)
$T = 0$

we find current profiles as shown in fig. 10.9.

The values of the residual surface currents theoretically found agree well with those deduced in section 6.10 for moderate winds in a layer of 0.5-1 m depth. It can be seen that a southerly wind would give rise to a pattern that is similar to a so-called anti-estuarine circulation. However, for more persistent southerly winds, this pattern after some time would cause an increase in the salinity gradient by obstructing the outflow of low salinity water. The resulting changes in the pressure distribution would more or less compensate the effect of the wind.

The circulation during calms shows very low current velocities. Apparently the wind is more important than the density distribution as a cause for the circulation pattern.

Furthermore, the theory of Hansen and Rattray serving as a basis for a classification system of estuaries will be treated in section 11.2.

It may be mentioned that in another study RATTRAY (1967) gives solutions for fjord-type circulation, in which the vertical variation of the vertical eddy coefficients A_v and K_v is taken into account. The circulation is confined to the upper layers.

This latter supposition appears not to be valid for the Ría de Arosa. So here we have not made a further comparison of Rattray's results with our findings, although a vertical variation of A_v and K_v otherwise gives a more realistic approach than a value that is independent of depth.

10.9 Application of the circulation model to the distribution of oxygen and nutrients

The circulation model can be applied to the distribution of oxygen and nutrients. Thus we may calculate the changes in the concentration of these substances due to the diffusive and advective transport processes.

For the upper and deep layers of the different segments all observations of the concentration of dissolved oxygen and of the nutrients (PO_4-P, NO_3-N and NO_2-N) have been averaged. The number of observations of these properties is much smaller than that of salinity and temperature. Therefore, in view of the relatively large variability of these properties, and especially of the nutrient-concentrations, variations in the mean values were found that appeared to be not representative, and that would have given quite senseless results if used with the circulation model. Therefore a smoothing procedure was applied by which the observations of adjoining segments were also taken into consideration, be it with half the weight of the observations of the segment under consideration. So, if C_k is the mean concentration in segment K, based on N_k observations, the concentration after smoothing is

$$\overline{C_k} = \frac{C_{k-1} + 2\,C_k + C_{k+1}}{N_{k-1} + 2\,N_k + N_{k+1}}$$

In fig. 10.10 the smoothed longitudinal distribution of the various concentrations is given. From these values the transport of oxygen, phosphate, nitrate and nitrite may be calculated, using the circulation model of table XVII. The results of these calculations are given in table XIX, giving the surplus or deficit of oxygen or nutrients in the different segments of the ria.

TABLE XIX.

segment	$\triangle O_2$ upper layer l/s	$\triangle O_2$ deep layer l/s	$\triangle PO_4$-P upper layer g/s	$\triangle PO_4$-P deep layer g/s	$\triangle NO_3$-N upper layer g/s	$\triangle NO_3$-N deep layer g/s	$\triangle NO_2$-N upper layer g/s	$\triangle NO_2$-N deep layer g/s
B	— 400	+ 20	+10.7	—3.0	+ 22.2	— 1.5	+2.3	—0.5
C	— 220	—135	+ 2.1	—1.5	+ 16.3	+ 8.3	+0.9	—0.1
D	—1200	—160	+10.8	—1.8	+ 42.8	+16.9	+2.4	+0.3
E	—1080	+ 40	+ 9.9	+0.6	+ 39.8	— 1.0	+2.9	+0.1
F	+ 110	— 30	+ 2.3	—4.4	+ 7.7	—24.6	+0.3	—1.2
Total	—2790	—265	+35.8	—10.1	+128.8	+ 1.9	+8.8	—1.4

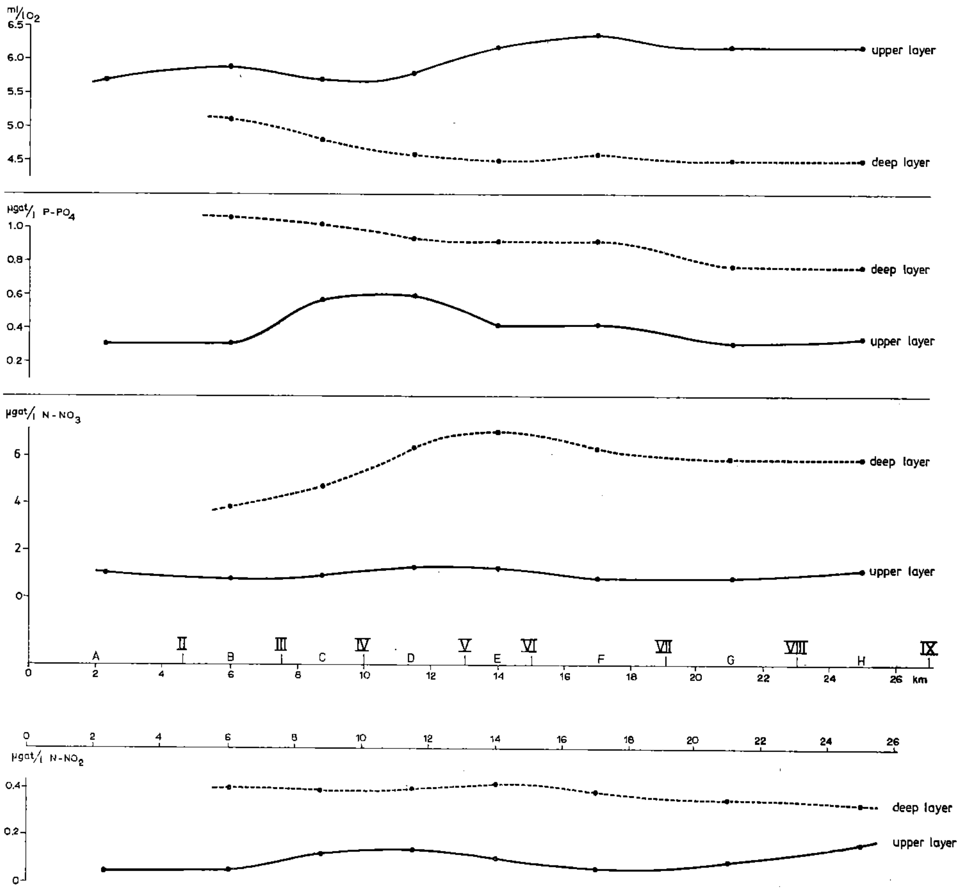


Fig. 10.10 Variation of concenration of nutrients along the axis of the ria.

10.10 Discussion of the results

The surplus and deficit of oxygen and nutrients by differences in transport as given in table XIX may be related with the biochemical processes that are taking place. Let us first consider the increase of phosphorus in the upper layers. Under stationary conditions the accumulation of phosphate can be expected to be balanced by photosynthetic assimilation which converts phosphate into living matter. Near-stationary conditions are likely except during periods of blooms or massive die-off of plankton.

Suppose the increase is completely balanced. Because 1 atom of phosphorus is used in the photosynthesis of 106 atoms of carbon (see REDFIELD, KETCHUM AND RICHARDS, 1963), the primary production of the different segments is found to be

segment	B	production	1.4 gC/m^2. day
	C		0.4
	D		1.9
	E		1.5
	F		0.4
mean value for whole area			1.2

These values are in reasonable agreement with the production rate estimated from the diurnal variation of dissolved oxygen at Punta Preguntoiro, and with the estimate of VIVES AND FRAGA (1961B) for a 20 m thick layer in the Ría de Vigo. They are also close to a value published by CURRIE (1958) for the production at a station 15 miles off Oporto, september 1955 (1.1 g/m^2 day). From these production figures we may estimate the amount of nitrogen utilized in the upper layers, as under normal conditions, that is if not a so-called deprivation or luxury feeding of phytoplankton occurs, 1 atom of phosphorus is used together with 16 atoms of nitrogen (see REDFIELD, KETCHUM AND RICHARDS, 1963). Now the surplus of nitrogen available in the form of nitrate and nitrite is, according to table XIX, smaller than the requirement that is calculated from the estimated production. The remaining nitrogen may be present in the form of ammonia, which according to VACCARO (1965) appears to be used preferentially in photosynthesis. Ammonia concentrations have not been measured during the summer, but they may attain significant values judging from studies in the Ría de Vigo (FRAGA, 1967).

Over the whole inner and central ria we find a net surplus by transport of dissolved nitrate and nitrite (advection and diffusion) of 137 g nitrogen per second. The amount of nitrogen utilized by photosynthesis with the assumptions mentioned above can be estimated to be 259 g nitrogen per second. So there is a difference, estimated to be about 122 g/s, which can be thought to be supplied by the transport of ammonia into the upper layers.

Attempting to calculate the balance of oxygen for the upper layers, using the earlier mentioned atomic relation for photosynthesis we come across the difficulty that the source for the nitrogen utilized for photosynthesis (NO_3, NO_2 or NH_4) influences the result. If the nitrogen is present in the form of nitrate the assimilation gives 4 free atoms of oxygen per atom N assimilated, while for nitrite 3 free atoms oxygen come available per atom nitrogen (VACCARO, 1965).

We find that for the inner and central ria advection and diffusion of dissolved oxygen accounts for a loss of 2800 l/s. Photosynthesis accounts for a gain of 3600 l/s. An estimate of the contribution of nitrogen assimilation to the oxygen balance gives something of the order of 400 l/s. So there is a difference of about 1200 l/s, and a flux of oxygen from the surface waters to the atmosphere has to be assumed in order to equalize the balance.

In the deep layers the situation is rather complicated. There is according to table

XIX a deficit of phosphate which, under stationary conditions may be balanced by production of phosphate through mineralization of dissolved or suspended organic matter or by supply of phosphate from the bottom sediments. An exchange with the bottom, either as supply to the bottom or to the water is not unlikely. For the bottom sediments of the Ría de Vigo FERNANDEZ DEL RIEGO (1958) found concentrations of organic and inorganic phosphorus between 0.05 and 0.025 % (by weight). This phosporus has been deposited with sedimentary material. Whether such a process of loss of phosphorus to the bottom is active during the summer is not certain, nor is it known in what form this phosphorus is deposited, or whether there is an extra oxygen requirement for oxydation of organic bottom material.

For a complete remineralization of plankton each atom of phosphorus requires 276 atoms of oxygen. However, it is likely that much of the organic material is already in some state of decomposition. According to MOTOHASHI AND MATSUDAIRA (1969) we may expect that for this material the oxygen requirement is relatively less.

A further point is that some oxygen is used or delivered by shifts in the ammonia-nitrite-nitrate system. Table XIX shows an overall surplus of nitrate and a deficit of nitrite by advective and mixing processes. Remineralization of organic matter is a source of anorganic nitrogen. As data on ammonia concentrations are lacking the estimate of the different terms of the nitrogen balance is even more difficult than for the phosphorus balance.

Summing up, there are too many unknown factors in the whole nutrient-oxygen system of the deep layers to allow a further investigation of the processes involved. Considering the different items in the balance it is doubtful whether conditions are close to the stationary situation.

Especially, it is likely that during the summer the concentration of oxygen in the deeper layers has the tendency to decrease.

10.11 The total nutrient budget in the inner and central ria

In an coastal embayment with an estuarine circulation conditions are favourable for an accumulation of nutrients (see e.g. KETCHUM, 1967). To investigate this point for the Ría de Arosa we have to compare the inflow and outflow of the nutrients in the different modes in which they may be present. A nutrient element as phosphorus is mainly present in the form of dissolved organic substances, as dissolved inorganic phosphate and in the form of suspended matter, including plankton.

The transport of inorganic phosphorus in and out of the inner and central ria follows from the data of table XXIII. The deep layer looses 10.1 g PO_4-P per second, while the upper layer gains 35.8 g/s. This gives a net supply of 25.7 g/s.

The supply or loss of dissolved organic phosphorus may be estimated from the surplus of dissolved organic carbon. The average concentrations of organic carbon in the upper and deep layers of the inner and central ria on the one hand and of the outer ria on the other hand are calculated from the data given in table XIV. To calculate these mean values stations 62/222, 63/58 and 63/60 were used, for the inner and central ria, and stations 63/55 and 63/58 for the outer ria.

inner and central ria	upper layer	1.2 mg/l organic C
	deep layer	0.75 mg/l
outer ria	upper layer	1.25 mg/l
	deep layer	0.9 mg/l

Combining the circulation data from table XVII with these figures a net outflow of about 700 g dissolved organic carbon per second is found. This figure, however, is only a rough estimate and should be regarded to give only an indication of the order of magnitude of this outflow. From this figure we may estimate the loss of dissolved organic phosphorus if the mean carbon/phosphorus ratio of the dissolved organic substances involved is known. DUURSMA (1961) found for the North Sea the carbon/ phosphorus atomic ratio to be usually about 250 : 1. If this result is also applicable to the substances present in the Ría de Arosa, (regardless of depth) we find a loss of dissolved organic phosphorus of about 7 g per second. Compared with the net surplus of 26.5 g/s of dissolved inorganic phosphorus this is of secondary importance.

The balance of suspended matter may give an estimate of the supply or loss of phosphorus in suspended form. The number of data on the concentration of suspended matter is sufficient to give an estimate of this part of the balance based on the values of the concentration in segments F and G. When making this estimate it is assumed that there is no important vertical variation of the concentration in either of the two layers. As stated in section 8.7 the concentrations close to the bottom show a tendency to be somewhat higher, but this effect does not seem to play an important part in this case. The mean values of the concentration (dry weight) are as given in the following table.

segment F	upper layer	8.5 mg/l
	deeper layer	7.4 mg/l
segment G	upper layer	9.3 mg/l
	deeper layer	7.8 mg/l

The loss of suspended matter calculated from these figures is about 2400 g/s. The inflow of suspended matter at the head of the ria, out of the river (which is estimated to be about 100 g/s), may be mainly sand or silt. About the inflow from the ocean we are not sure, but supposing that this is mainly organic material in accordance with the results of CAPIN-MONTEGUT (1972) for the sea area west of Spain and that erosion within the ria may be neglected, a quantity between 2100 and 2200 g of organic particles is lost from the inner and central ria each second.

Apart from this, there is also the loss by sedimentation, but this is thought to be small (see par. 8.8). Supposing all this suspended matter would consist of phytoplankton which, according to STRICKLAND (1965) contains roughly 1 % of phosphorus, then there would be a loss of about 22 g phosphorus per second. This is about equal to the estimated surplus of inorganic phosphorus (26.5 g/s). For a suspension of other organic particles (mainly seston) the equivalent loss of phosphorus is difficult to assess but a figure of the same order of magnitude is likely with a view to the results of CAPIN-MONTEGUT (1972). We may conclude that, because of the rather inaccurate estimates, we can only say that the total budget of phosphorus is close to a complete balance. A process of accumulation of nutrients that nevertheless may be acting apparently cannot be detected within the limits of accuracy that could be attained.

11. Comparison with other estuaries and embayments

11.1 Estuarine classification

The Ría de Arosa meets the definition of an estuary as given by PRITCHARD (1967), that is: 'An estuary is a semi-enclosed coastal body of water which has a free connection with the open sea and within which sea water is measurably diluted with fresh water derived from land drainage'.

Other definitions include so-called 'negative estuaries' or 'neutral estuaries'. With 'negative estuaries' (or 'inverse estuaries') are meant embayments where the salinity is higher than in the adjoining sea area. The circulation pattern often occurring in such embayments is the inverse of the pattern often observed in true estuaries and is called 'anti-estuarine circulation'. 'Neutral estuaries' are embayments in which neither evaporation nor freshwater inflow dominate (see EMERY AMD STEVENSON, 1957).

Although the run-off is low in summer, the influence of the river water can be clearly observed in the Ría de Arosa. So-called 'neutral' conditions were observed in a separate part of the ria (section 6.17), but this does not influence the circulation pattern as a whole.

If, following Pritchard, we only consider the so-called 'positiev' estuaries, we may attempt to classify them according to different points of view. From the geomorphological point of view there are, according to Pritchard, four primary subdivisions:

1. drowned river valleys
2. fjord-type estuaries
3. bar-built estuaries
4. estuaries produced by tectonic processes.

As regards the present study the geomorphological characteristics are mainly of interest in so far as they are related to the circulation pattern. Pritchard notes that a drowned river valley often embraces an upstream region in which only fresh water is present, and a downstream region, containing more or less brackish water, which latter part is only considered an estuary in accordance with the definition. Fjords are described as coastal indentures that have been gouged out by glaciers. Frequently they have a shallow sill at their mouth. Such sills are generally important with respect to the circulation. Bar-built estuaries are coastal areas enclosed by sand spits. Often they are relatively shallow and the tide in the interior may be reduced, compared with the tide outside the bar. Therefore wind effects may be relatively more important for the circulation than for other estuaries. Many sea areas enclosed by an outer bar have a high evaporation and many so-called 'negative estuaries' belong to this geomorphological class.

Estuaries produced by tectonic processes can be very different in their characteristics and no more or less typical circulation pattern can be assigned to this class.

The west-Galician rias have been considered the prototypes for drowned river valley in a submerged coast. Although the processes that have formed these rias are rather complex (PANNEKOEK, 1966B, 1970), they best can be placed into the first category. There is, however, some morphological resemblance with the fjord type.

For comparison of the Rías Bajas with fjords we may refer to FERNANDEZ DEL RIEGO (1958). The main difference is that deep basins behind a relatively shallow sill as usually occur in fjords are not present in the Rías Bajas. As seen in section 1.5 the Ría de Arosa also has a distinct sill (the 'central rise') and a deeper basin inside, but the dimensions of these features are so much different from those found in the fjords, that their oceanographic significance is not similar.

Sand-spits occur in the Ría de Arosa, and we may ask ourselves whether the southeastern part has not some characteristics of a bar-built estuary. As, however, the main emphasis of our work has been with the main channel of the ria we will consider this point no further.

In connection with this study a classification based on the physical processes that determine circulation and mixing is of more interest. Here wind, tide and river inflow are the basic processes that are decisive, besides the topography. PRITCHARD (1955) has given a classification that is very useful. Classifications along the same lines have also been given by CAMERON AND PRITCHARD (1963) and by BOWDEN (1967).

We will follow the classification as given by Bowden. There are the following categories:

1. Salt wedge estuary. It is essential that possible tidal movements are small. River flow is dominant. The sharp interface between the river water and the sea water is nearly horizontal. The water in the deep layer has practically no net inward motion. In both the surface and the lower layer the salinities are practically constant.

2. Two-layer flow with entrainment of salt water into the fresh or brackish surface layer. Here also the river flow is important, but dynamical instability of the boundary between the two layers causes breaking internal waves which results in an upward motion of deeper water. There is a small net inward motion of the deep layer to compensate for this upward flow. The salinity of the upper layer increases downstream, the salinity in the lower layer is constant.

3. Two-layer flow with vertical mixing. Here vertical mixing between the upper and lower layer is caused by tidal motion. There is an inflow along the bottom as in the previous case, but the salinity in the deep layer decreases towards the head of the estuary.

4. Vertically homogeneous estuaries, where vertical mixing takes place under influence of strong tidal currents. If the estuary is wide, lateral differences may occur and there may be a net inflow of sea water on one side and a net outflow of river water on the other side (class 4a). If the estuary is relatively narrow, the inward transport of sea water occurs by longitudinal mixing (class 4b).

Fjords sometimes are regarded as a separate class, because of the presence of a deep, more or less stagnant layer below sill depth. But as regards their circulation above sill depth they commonly may be classified in group 2. HANSEN AND RATTRAY (1966) give a classification scheme to which we will refer in the next section. In their scheme stratification and circulation are the two characteristic properties.

In practice it is very difficult to find straightforward examples of the types mentioned above. BOWDEN (1963) has demonstrated that even for an apparently homogeneous estuary as the Mersey Narrows shear effects play an important part in the quasi-turbulent diffusion of sea water upstream. Furthermore in this classification scheme the special influence of the wind is not considered. In the Ría de Arosa this effect is at least of equal importance as the effect of the density differences. Leaving aside this point for the moment we find that the inner and central ria could best be classified as an intermediate form between type 2 and 3 estuaries, while the lower course of the Ulla river exhibits characteristics intermediate between type 3 and type 4b. But as lateral differences also occur in the ria, especially in the outer ria, even some characteristics of type 4a estuaries are found.

As regards the influence of the wind, we cannot, strictly spoken, use a classification scheme that is based on the effect of density differences only. In practice, however, there are so many similarities in the circulation pattern that characterizes the type 2 and 3 circulations and a wind-driven circulation, that the scheme given here is not without sense. We may refer to BRONGERSMA-SANDERS (1965) who uses the term 'estuarinelike circulation' for cases where wind effects are dominant. We should note that there are secondary effects of the wind that in some way may counteract the density-driven circulation. The wind-induced upwelling at the head of the estuary may lower the surface temperature and so reduce density differences.

Although very useful the classification schemes as given here are rather qualitative. There is some need for a more quantitative approach. In the following paragraph we shall give some attention to this aspect.

11.2 Estuarine classification according to Hansen and Rattray

The theory for estuarine circulation mentioned in section 10.8 has served as a basis for a classification system developed by these authors. (HANSEN AND RATTRAY, 1966).

Although some discrepancies between this theory and the situation in the Ría de Arosa have been described earlier, these differences are not considered to be so serious as to impede a useful application of this scheme to the present case.

Two parameters are used in this classification: a circulation parameter ($\frac{u_s}{U_f}$ or the ratio of the residual surface current to the mean freshwater velocity through a certain section) and a stratification parameter ($\frac{\triangle S}{S_0}$ or the ratio of the top-to-bottom salinity difference to the mean salinity over a section). For the inner part of the Ría de Arosa (section C) we find $\frac{u_s}{U_f} = 10^3$ and $\frac{\triangle S}{S_0} = 2 \times 10^{-2}$. This places the Ría de Arosa in class 3a of Hansen and Rattray's system, a class which also contains the Strait of Juan de Fuca (Pacific coast of Canada), and which is the small-stratification counterpart of the class in which most fjord-type estuaries are found.

In their study of estuarine classification Hansen and Rattray define two so-called 'bulk parameters', which are parameters that only depend on river flow, tides and geomorphology. Stratification and circulation, given by $\frac{\triangle S}{S_0}$ and $\frac{u_s}{U_f}$ respectively are then empirically related with these bulk parameters; once established, such relations may be used to investigate the effect of for instance variations of river flow on the circulation pattern. Apparently the wind effect is left out of consideration in this case.

The bulk parameters defined are the 'densimetric Froude number' $\boldsymbol{F}_m$ and $\boldsymbol{P}$, the ratio between mean fresh water velocity U_f and the root mean square of the tidal current speed U_t

$$\boldsymbol{F}_m = \frac{U_f}{\sqrt{gd(\rho_{\text{sea}} - \rho_{\text{river}})/\rho}}$$

and

$$\boldsymbol{P} = \frac{U_f}{U_t}$$

Their values for the Ría de Arosa can be determined and compared with the circulation parameters as defined by Hansen and Rattray and established in section 10.8. We find $\boldsymbol{F}_m = 1.0 \times 10^{-4}$ and $\boldsymbol{P} = 0.8 \times 10^{-3}$. The empirical relations of Hansen and Rattray between these bulk parameters and the circulation parameters don't apply to our case. One of the reasons for this poor agreement may be that the effect of the wind is disregarded, which effect, as we have seen, is very important in the summer situation. Another reason pointed out by Hansen and Rattray may be the relative importance of boundaries on the velocity profile if the density differences are small, a situation we have in our case. However, in winter the influence of the wind is probably less important for the average conditions, and we may hope that the empirical relations mentioned above give rather reliable indications for the expected circulation pattern. The values of $\boldsymbol{F}_m$ and $\boldsymbol{P}$ for the winter are estimated to be $\boldsymbol{F}_m = 0.9 \times 10^{-3}$ and $\boldsymbol{P} = 7.5 \times 10^{-3}$. This brings the circulation in winter in class 2a which also contains the Mersey Narrows. Whether this is in accordance with reality cannot be said, as our winter observations were made under exceptional circumstances.

11.3 Comparison with the Ría de Vigo

The only ria of the Rías Bajas of which the physical and chemical oceanography has been investigated up till now is the Ría de Vigo. SAIZ ET AL. (1957) have made a thorough study of the properties of the water in this ria in connection with the meteorological and hydrological situation, and presented a theoretical model of the circulation. In a later study SAIZ ET AL. (1961) give further information on the situation in the Ensenada de San Simón, the innermost part of the Ría de Vigo, which because of the small depth and large influence of fresh-water run-off exhibits considerable variations in salinity and temperature. A third study (ANADÓN ET AL., 1971) gives details on the annual variations of the temperature and salinity structure.

These studies show a situation that is very similar to that of the Ría de Arosa. As they cover not only the summer, but the whole year, they are extremely important for an appreciation of the situation in the Ría de Arosa at other times of the year. In this connection it is interesting to note that in the Ría de Vigo a more or less marked stratification appears to be the normal case in all times of the year. Only in some exceptional cases in the winter temperature and salinity are about the same for the surface and the deep layers.

The authors of the studies mentioned above given evaluations of the circulation processes that are different from those that have been given in the present study for the Ría de Arosa. The flushing model for the Ría de Vigo as given by SAIZ ET AL. (1957) is based on a modification of the 'tidal prism' concept as developed by KETCHUM (1951). This concept assumes among other things a thorough vertical mixing of the water during certain phases of the tidal cycle. As was shown in the foregoing chapters the water in the Ría de Arosa is stratified at all phases of the tide and the same appears to be the case for the Ría de Vigo. In an application of his theory to Alberni Inlet, a typical stratified fjord, Ketchum only considered the layer comprising the upper 30 feet. Probably this would have been a more appropriate procedure for the case of the Ría de Vigo. It is true that Saiz et al. in their study have recognized the problem imposed by the stratification of the waters, but they have nervertheless used a circulation model in which vertical stratification and vertical variations of the water transport are neglected. It is apparent that such a circulation model will fail if it is used for the calculation of budgets of dissolved substances with a marked vertical variation, such as nutrients.

The influence of the wind, which is regarded as one of the important aspects of the situation in the Ría de Arosa is specially noted by SAIZ ET AL. (1961) for the shallow Ensenada de San Simón, but these authors were unable to separate the effect of the wind on the hydrographical situation from the effects of other environmental factors.

11.4 Other rias of the Galician coast

No systematic oceanographic studies are known of the two other 'Rías Bajas', the Ría de Muros y Noya and the Ría de Pontevedra. There are only some series of oceanographic observations, made by the Spanish research vessel 'Xauen'. The first series of observations, made in the Ría de Muros y Noya on September 14 and 15, 1954 (INSTITUTO ESPAÑOL DE OCEANOGRAFÍA, 1961) shows in the deeper central and outer region a stratification similar to that in the Ría de Arosa and an only slightly depressed salinity together with comparatively low temperatures at the surface in the inner parts. Two other series of observations on September 20 and 21, 1949 in the Ría de Muros y Noya and on October 7 and 8, 1949 in the Ría de Pontevedra were made during the same cruise as the observations in the Ría de Arosa mentioned in section 6.21 (INSTITUTO ESPAÑOL DE OCEANOGRAFÍA, 1955). As was the case with these latter observations the data for both the Ría de Muros y Noya and the Ría de Pontevedra show only a slight stratification. For the Ría de Arosa this was taken as an indication of a change in the oceanographic situation prevailing in summer. They apparently are not representative for the normal summer conditions. We may expect that, because of their general resemblance in morphology and in the meteorological and hydrological conditions the two-layered circulation that was found during summer in the Ría de Arosa and that also exists in the Ría de Vigo is a normal situation in all four Rías Bajas.

Besides the four Rías Bajas there is a large number of smaller rias along the west and north coasts of Galicia. However, no studies of the circulation of these estuaries are known, so we can only derive our information from other studies in which the aspect of circulation has not been given special emphasis.

FISCHER-PIETTE AND SEOANE-CAMBA (1963) give some data of surface temperature and salinity of the Ría de Camarinas, a small ria just north of Cape Finisterre. Although the observations were made in October, the surface temperatures are equal or even higher than the summer values of the Ría de Arosa. The depth of the area that was studied is less than 10 m and this makes the occurrence of such a pronounced two-layer circulation not very probable. Only the outer part of the ria is deeper, but here no observations were made.

As regards the rias on the north coast of Spain some temperature and salinity data are found for the Ría del Barquero in an ecological study of FISCHER-PIETTE AND

SEOANE-CAMBA (1962). These data from the inner part of the estuary near the mouth of the river Sor show a situation that is typical for a river estuary where a mixing of freshwater and sea water takes place under influence of the tide. Judging from studies of HERNÁNDEZ PACHECO AND ASENSIO AMOR (1963, 1965 end 1966) on the sedimentology of the Ría del Eo and the Ría de Guernica these rias of the north coast are furthermore influenced by the tide and the coastal currents induced by sea and swell from the north.

11.5 Conditions in some other comparable embayments

The combination of a low river discharge and a rather steady wind, as found during the summer in the Rías Bajas is typical for coastal areas in the subtropics. These are the regions where embayments with circulation patterns similar to the rias are most likely to be found.

We find a good example in the Gulf of Cariaco on the coast of Venezuela. This gulf has been investigated by several authors (RICHARDS, 1960; GADE, 1961A, 1961B). The Gulf of Cariaco is about 60 km long, which is about twice the length of the rias. There is a sill of about 60 m depth, separating the inner basin with depths over 90 m from the ocean. The river discharge in the gulf is negligibly small, especially in the dry season (winter and spring) and evaporation exceeds precipitation for most of the time. This might give an anti-estuarine circulation if it were not for the wind. The wind is predominantly easterly, towards the ocean, and shows a diurnal variation of windforce. Only in summer the winds are less persistent, and at that time of the year winds from other directions or calms occur. Upwelling takes place in the sea area outside the gulf. A special point of difference with the Rías Bajas is the small tidal range (0.5 m).

The circulation in the Gulf of Cariaco is mainly winddriven although GADE (1961B) gives reasons to believe that during calms an anti-estuarine circulation may occur. During periods of increased stability, caused by a weakening of this circulation pattern and by higher surface temperatures at the entrance of the gulf the water below sill depth becomes separated from the overlying water. Especially in this situation anoxic conditions occur and hydrogen sulflide develops in these deep layers.

In section 11.2 it was found that, according to the classification scheme of Hansen and Rattray, the circulation in the Ría de Arosa during the summer is in the same class as the circulation found in the Strait of Juan de Fuca. Here we may give some characteristics of this sea area as found in literature.

The Strait of Juan de Fuca is about 20 km wide and some 200 m deep. Its surface waters are diluted by less saline water, the main source being the water from the Strait of Georgia, which itself receives fresh water from the Fraser river. Before entering the Strait of Juan de Fuca the Strait of Georgia water mixes with deep water from the Juan de Fuca Strait in the narrow and shallow channels connecting these two straits (see WALDICHUK, 1957). This mixture flows towards the ocean over the inflowing deep water, as is demonstrated by current measurements made by HERLINVEAUX (1954B). According to these measurements tidal currents are strongly developed. The wind at large is predominantly northwest in summer and south or southeast in winter, but in the coastal zone with its complicated topography deviations occur, so here the contribution of the wind cannot be made out. Judging from data of HERLINVEAUX (1954A) the effect of the wind has a transient character. It therefore appears that the tidal mixing is mainly responsible for the movement of water from deep into surface layers and not the wind. Anoxic conditions have been found in typical fjords (with shallow sills) along the Strait of Georgia-Strait of Juan de Fuca system (RICHARDS 1965), but have not been reported in these straits themselves.

11.6 Concluding remarks

Concluding we may note that the conditions as found in the Ría de Arosa in summer and probably also present in the other Rías Bajas have no well-known couterpart elsewhere. They are somewhat intermediate between the conditions found in some fjords with a mainly density-driven circulation and those found in the Gulf of Cariaco with its wind-driven circulation. That no anoxic conditions occur in the Rías Bajas (at least as far as is known), while they are found in both other types of embayments, is probably due to the following reasons:

1. relatively small length, which makes possible a rather fast renewal of water (section 10.6);
2. small difference between the depths of the sill and the basin behind it, which helps to prevent the occurrence of a deep stagnant layer;
3. tidal motion, which creates horizontal and vertical mixing;
4. rather brief periods in which conditions occur that are favourable for accumulation of nutrients and consequent increased production of organic material.

As is suggested in section 10.10 the oxygen budget of the deep layers would have a deficit. We may speculate under what conditions this might lead to near-zero oxygen values. This might be the case if renewal of water would be slowed down and if organic material would further accumulate. Under natural conditions (i.e. neglecting pollution by organic wastes) this could happen if, at the end of a period with a two-layered summer circulation, the rate of circulation diminishes because of slackening of northerly winds and more regular moderate south-westerly winds (not strong enough to cause important vertical turbulence) or calms. If under these conditions the density-driven circulation is not strongly enhanced by increased river flow, the rate of renewal might become sufficiently low to give, after some time, a significant lowering of oxygen values.

It may be noted that such conditions would also give an increased stability of the water because of the diminished upwelling of cold water. It may also be noted that, according to MARGALEF (1956), this is the situation which may be favourable for plankton bloom, locally known as 'purga de mar'. It would be interesting to study the conditions in late summer and early autumn to investigate this effect.

ANNEX

ANNEX 1

Some data on the annual meteorological variation.

The following data have been derived from the following sources:

(1) Friedemann, 1912
(2) Markgraf and Bintig, 1956
(3) Markgraf and Bintig, 1954
(4) Meteorological Office, 1944
(5) Semmelhack, 1910
(see the list of references)

Wind Table A1

Annual variation of wind direction.

Source: (4)
Pontevedra: obs. at 08 and 16 GMT.

Table A2

Annual variation of the frequency of winds > 8 Bft.

Source: (2) and (3)

Temperature Figure A1.1

Mean monthly temperatures at sea, for Pontevedra and Santiago de Compostela.

Sources: (4) and (5)

Humidity Figure A1.2

Atmospheric humidity at Pontevedra.

Fog Table A3

Mean number of days with fog at Pontevedra and frequency of visibility less than ½ n.mile, sea area 42°-44° N and 8°-10° W.

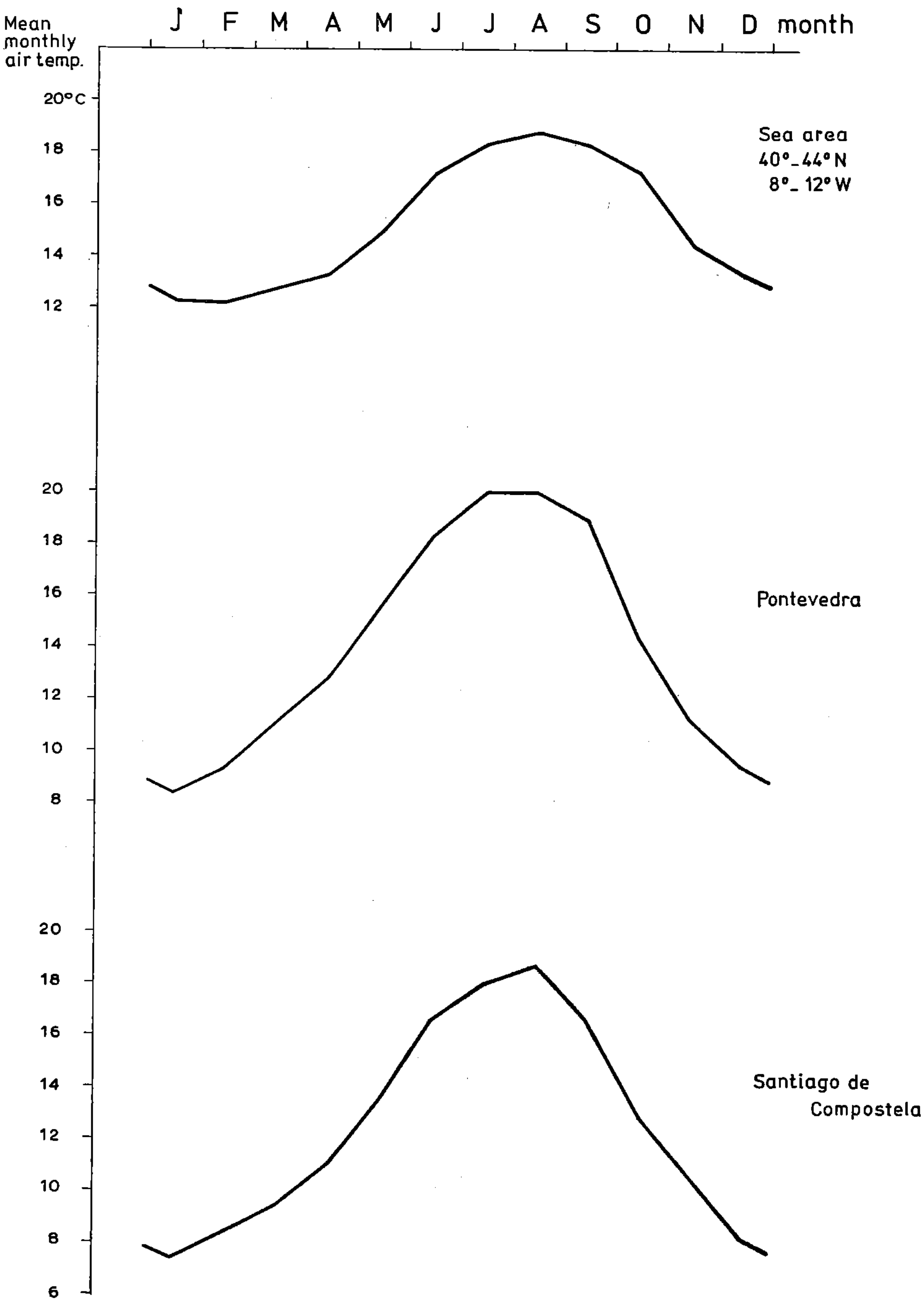
Source: (4)

TABLE A1. *Annual variation of wind direction (in % of obs. per month)*

Sea area 40°-43° N, 8°-12° W (1887-1899 and 1921-1938)										Pontevedra (10 year period, years not indicated)								
month	N	NE	E	SE	S	SW	W	NW	calm	N	NE	E	SE	S	SW	W	NW	calm
J	16	17	8	6	14	13	11	10	3	26	10	3	13	26	6	6	10	—
F	22	16	5	4	11	12	13	14	2	32	14	4	11	21	7	4	7	—
M	22	14	8	6	10	12	9	13	3	28	17	0	6	17	10	6	16	—
A	28	9	3	3	9	12	16	18	2	42	17	0	3	8	12	6	12	—
M	36	9	3	3	6	9	13	18	3	30	10	0	3	26	10	6	15	—
J	39	12	2	1	5	12	9	15	3	44	17	0	3	10	3	3	20	—
J	42	9	1	1	4	9	12	21	2	36	19	0	0	16	13	6	10	—
A	46	10	1	2	5	9	10	13	3	49	14	0	3	19	6	3	6	—
S	35	14	3	4	7	10	9	13	4	40	20	3	0	17	10	3	7	—
O	24	14	6	6	14	11	10	11	2	39	23	3	6	13	3	3	10	—
N	18	16	11	8	11	10	13	12	1	27	26	0	7	13	3	7	17	—
D	17	14	7	8	13	16	12	10	3	23	23	3	3	26	6	3	13	—

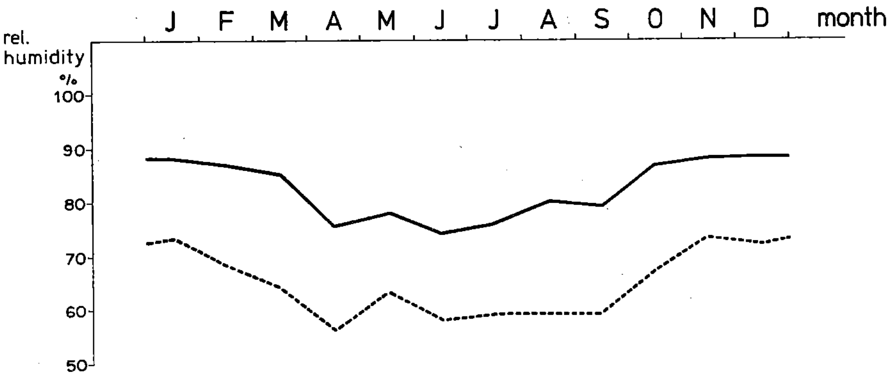
TABLE A2. *Annual variation of the frequency of winds with windforce 8 Bft. ore more (in % of obs. per month)*

Sea area 41°-43° N, 9°-11° W, years 1906-1939								
month	N	NE	E	SE	S	SW	W	NW
J	0.5	0.8	0.2	0.2	1.3	1.4	1.3	1.0
F	1.1	0.6	0.2	0.1	1.0	1.6	1.7	1.5
M	0.7	0.7	0.1	0.1	0.5	0.8	1.0	1.0
A	1.0	1.4	0.2	0.1	0.4	0.4	0.5	0.6
M	0.6	0.5	0.0	0.0	0.1	0.2	0.1	0.0
J	0.7	0.3	—	—	0.0	0.1	0.2	0.3
J	1.0	0.4	0.0	—	0.1	0.1	—	0.0
A	0.5	0.4	—	—	—	—	—	0.0
S	0.3	0.2	0.0	0.0	0.1	0.1	0.0	—
O	0.6	0.3	0.1	0.2	0.6	0.7	0.6	0.4
N	0.3	0.3	0.1	0.2	0.9	0.8	0.5	0.6
D	0.8	0.4	0.2	0.2	1.3	1.5	1.1	0.8



Mean monthly air temp.
J F M A M J J A S O N D month
20°C
18
16
14
12
Sea area 40°–44° N 8°–12° W
20
18
16
14
12
10
8
Pontevedra
20
18
16
14
12
10
8
6
Santiago de Compostela

Figuur A1.1



Pontevedra, relative humidity
—— 08 m. GMT.
-------- 16 m GMT.
(from Met. office, ...)

Figuur A 1.2

TABLE A3. *Number of days with fog at Pontevedra (average over 33 year period) and frequency of visibility less than ½ n. mile*

	J	F	M	A	M	J	J	A	S	O	N	D
nr. of days, Pontevedra	1	1	0.4	0.1	0.4	1	1	2	1	1	1	1
frequency, sea area	1	0	1	0.6	1	2	2	4	5	1	0.3	0

ANNEX 2

Some data on the diurnal meteorological variation in summer.

The following data are the results of the observations at Punta Preguntoiro, carried out during the three summer campaigns.

Wind Table A4

Diurnal variation of the wind at Punta Preguntoiro Wind speeds are expressed in wind forces of the Beaufort scale (international scale 1946), either from direct estimates or from conversion of instrumental observations.

Temperature Figure A2.1

Diurnal variation of air temperature at Punta Preguntoiro.

Humidity Figure A2.2

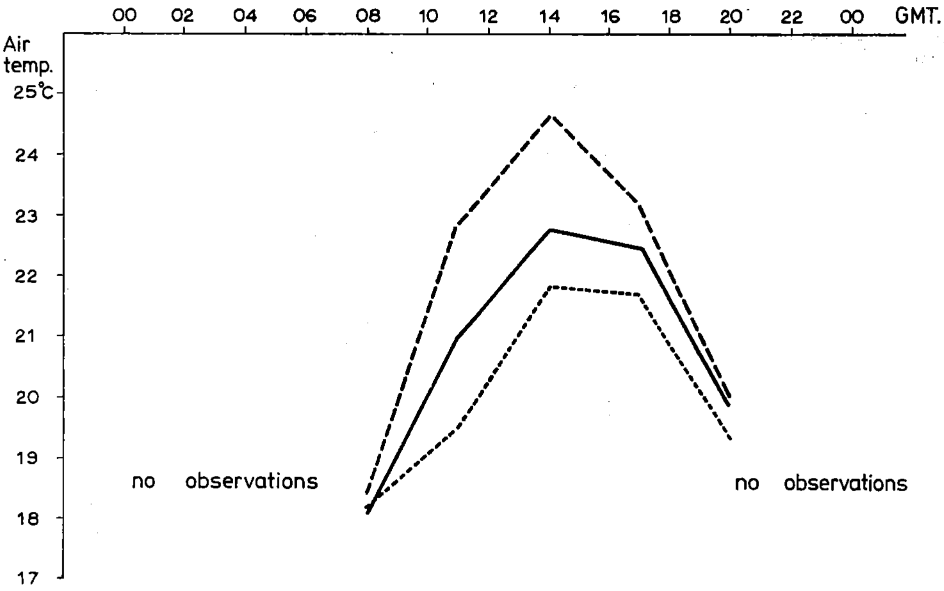
Clouds Table A5

Cloudiness at Punta Preguntoiro. Diurnal variation for all conditions and for Northerly and Southwesterly winds.

	08	11	14	17	20 GMT
all observations	4.2	4.0	3.3	3.1	3.6
wind NW-NE	2.4	1.8	1.9	2.3	2.1
wind S-W	8.2	6.2	4.2	3.8	5.5

TABLE A4. *Diurnal wind variation, observed at Punta Preguntoiro, July and August %*

time (GMT)	Wind force	N	NE	E	SE	S	SW	W	NW	calm
08 hr	1	4.9	0.7			1.4	2.1	2.1		22.2
(72 obs.)	2	6.2	0.7			2.1	2.8	0.7	1.4	
	3	7.0	2.1		0.7	2.8	2.8	0.7	0.7	
	4	14.6							0.7	
	5	12.5	4.9	0.7						
	6	2.8								
11 hr	1	3.0					0.8	0.8		1.5
(66 obs.)	2	0.8				0.8	3.8	6.8	3.0	
	3	7.6	3.0			3.0	6.1	14.4	3.8	
	4	10.6	5.3	0.8		3.8	3.8	3.0		
	5	11.3	2.3							
	6	1.5								
14 hr	1					2.1	1.4	1.4	0.7	2.8
(72 obs.)	2	1.4					4.9	7.0	0.7	
	3	11.1	1.4			4.9	9.7	5.6	2.1	
	4	8.3	2.8			0.7	7.6	11.1		
	5	6.2	2.8	0.7			2.1	0.7		
	6									
17 hr	1						1.5	1.5		1.5
(66 obs.)	2						6.1	3.0		
	3	7.6				2.3	16.0	7.6	1.5	
	4	17.4	1.5				6.8	6.8	2.3	
	5	13.7	1.5			0.8	0.8			
	6									
20 hr	1					3.1	8.3	5.2	4.2	12.5
(48 obs.)	2						5.2	5.2	2.1	
	3	6.2					7.3	5.2		
	4	14.6	2.1							
	5	17.7							1.0	
	6									



Punta Preguntoiro

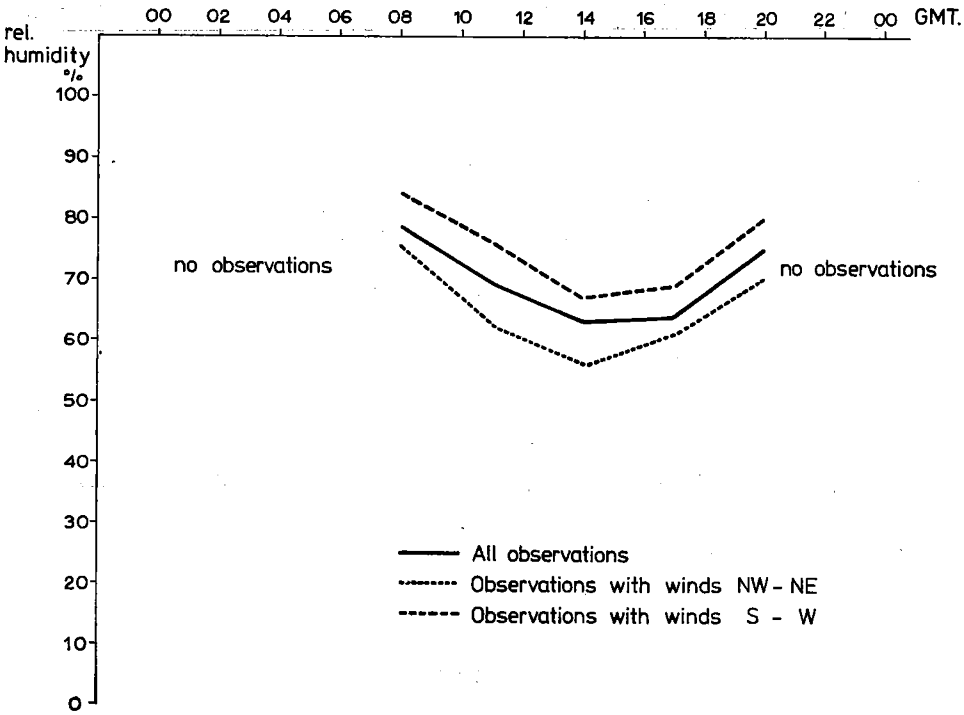
diurnal variation. of air temperature (july, august)

—— All observations

– – – – Observations with winds NW - NE

-------- Observations with winds S - W

Figuur A 2.1


rel. humidity %
100
90
80
70
60
50
40
30
20
10
0
00 02 04 06 08 10 12 14 16 18 20 22 00 GMT.
no observations
no observations
All observations
Observations with winds NW - NE
Observations with winds S - W

Figuur A 2.2

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Corrigenda

p. 18	fig. 1.2 station indicated A should be C station indicated C should be A
p. 34	3rd and 4th line of ch. 3.1. Delete words "for ... annex 1"
p. 35	lines 18 and 19 from top. Delete words "whereas ... 1900"
p. 37	7th line from bottom "June" should be "July"
p. 39	ch. 3.5, first formula $\frac{0,622}{p}$ to be shifted downwards, as a factor after D_w and not as an exponent
p. 39	7th line from bottom. After "been" add "given"
p. 41	Text of fig. 3.2. Insert "rate" between "evaporation" and "calculated"
p. 58	16th line of ch. 5.1 "acces" should be "access"
p. 58	last line of ch. 5.1. Insert "accurately" between "cannot" and "be"
p. 71	table VII. Frequencies in %
p. 72	8th line from bottom "consits" should be "consists"
p. 77	15th line from top "Uma" should be "Umia"
p. 82	5th line from bottom "hoistring" should be "hoisting"
p. 97	9th line of ch. 6.13 "durong" should be "during"
p. 127	5th line from top. Delete words "50 m deep"
p. 138	15th line from top "of" should be "if"
p. 143	3rd line from top, delete word "and"
p. 156	8th line of ch. 10.3, "0.98" should be "9.8 m^2/s^2 or 0.98"
p. 159	13th line of ch. 10.5 "Section 11.4" should be "section 10.4"
p. 160	7th line from top "... estimate. For ..." should be "... estimate, for ...".
p. 164	formula on 2nd line from top, 4th term D_i should be D_i'"
p. 165	table XVII. In top line K_v should be K_z
p. 168	17th line from top, insert "but" between "important" and "is"
p. 181	last line "primary" should be "net"
p. 190	14th line from bottom "given" should be "give"
p. 193	2nd line of ch. 11.6 "couterpart" should be "counterpart"

Van de reeks MEDEDELINGEN EN VERHANDELINGEN *zijn bij het Staatsdrukkerij- en Uitgeverijbedrijf nog verkrijgbaar de volgende nummers:*

23, 25, 27, 29*b*, 30, 31, 34*b*, 35, 36, 37, 38, 39, 40, 42, 43, 44, 45, 46, 47, 48, 50, 51, 52, 53, 54, 55, 56, 57, 59.

alsmede

60. C. Kramer, J. J. Post en J. P. M. Woudenberg. Nauwkeurigheid en betrouwbaarheid van temperatuur- en vochtigheidsbepalingen in buitenlucht met behulp van kwikthermometers, 1954. (60 blz. met 11 fig.) 3,60

62. C. Levert. Regens. Een statistische studie. 1954. (246 blz. met 67 fig. en 143 tab.). 10,30

63. P. Groen. On the behaviour of gravity waves in a turbulent medium, with application to the decay and apparent period increase of swell. 1954. (23 blz.). 1,55

64. H. M. de Jong. Theoretical aspects of aeronavigation and its application in aviation meteorology. 1956. (124 blz. met 80 fig., 9 krt. en 3 tab.) 4,60

65. J. G. J. Scholte. On seismic waves in a spherical earth. 1956. (55 blz. met 24 fig.). 5,15

66. G. Verploegh. The equivalent velocities for the Beaufort estimates of the wind force at sea. 1956. (38 blz. met 17 tab.) 1,80

67. G. Verploegh. Klimatologische gegevens van de Nederlandse lichtschepen over de periode 1910—1940.

Deel I: Stormstatistieken. — Climatological data of the Netherlands light-vessels over the period 1910—1940. P. I: Statistics of gales. 1956. (68 blz. met tabellen) 3,60

Deel II: Luchtdruk en wind; zeegang. — Climatological data of the Netherlands light-vessels over the period 1910—1940. P. II: Air pressure and wind; state of the sea. 1958. (91 blz. met tabellen.) 7,70

Deel III: Temperaturen en hydrometeoren; onweer. — Climatological data of the Netherlands light-vessels over the period 1910—1940. P. III: temperatures and hydrometeors; thunderstorms. 1959. (146 blz. met tabellen.). 8,25

68. F. H. Schmidt. On the diffusion of stack gases in the atmosphere. 1957. (60 blz., 12 fig. en tab.). 5,15

69. H. P. Berlage. Fluctuations of the general atmospheric circulation of more than one year; their nature and prognostic values. 1957.. 7,70

70. C. Kramer. Berekening van de gemiddelde grootte van de verdamping voor verschillende delen van Nederland volgens de methode van Penman. 1957. (85 blz., fig. en tab.) 7,20

71. H. C. Bijvoet. A new overlay for the determination of the surface wind over sea from surface weather charts. 1957. (35 blz., fig. en tab.). 2,60

72. J. G. J. Scholte. Rayleigh waves in isotropic and anisotropic elastic media. 1958. (43 blz., fig. en tab.) 3,10

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