

# YALE PEABODY MUSEUM

P.O. BOX 208118 | NEW HAVEN CT 06520-8118 USA | PEABODY.YALE. EDU

## JOURNAL OF MARINE RESEARCH

The *Journal of Marine Research*, one of the oldest journals in American marine science, published important peer-reviewed original research on a broad array of topics in physical, biological, and chemical oceanography vital to the academic oceanographic community in the long and rich tradition of the Sears Foundation for Marine Research at Yale University.

An archive of all issues from 1937 to 2021 (Volume 1–79) are available through EliScholar, a digital platform for scholarly publishing provided by Yale University Library at <https://elischolar.library.yale.edu/>.

Requests for permission to clear rights for use of this content should be directed to the authors, their estates, or other representatives. The *Journal of Marine Research* has no contact information beyond the affiliations listed in the published articles. We ask that you provide attribution to the *Journal of Marine Research*.

Yale University provides access to these materials for educational and research purposes only. Copyright or other proprietary rights to content contained in this document may be held by individuals or entities other than, or in addition to, Yale University. You are solely responsible for determining the ownership of the copyright, and for obtaining permission for your intended use. Yale University makes no warranty that your distribution, reproduction, or other use of these materials will not infringe the rights of third parties.



This work is licensed under a Creative Commons Attribution-NonCommercial-ShareAlike 4.0 International License.  
<https://creativecommons.org/licenses/by-nc-sa/4.0/>



# **Some historical, theoretical and applied aspects of quantitative water mass analysis**

by **Matthias Tomczak**<sup>1</sup>

## ABSTRACT

The concept of water masses is reviewed from the point of view of quantitative water mass analysis. A theoretical framework is presented which describes the life history of water masses in terms of formation, consolidation, aging and decay. Water masses are described as physical entities and compared with their atmospheric counterparts (air masses).

The classical temperature-salinity diagram is expanded into the mathematical concept of water types in an  $n$ -dimensional parameter space. Water types and their standard deviations are introduced as the foundation for quantitative water mass analysis. The relationship between parameter space and physical space is established through the definition of water type density. Mode Waters are discussed as regions in physical space with a minimum in water type density. Some unresolved issues of the structure of the oceanic thermocline are discussed in this context.

The definition of water masses is extended to include water masses in the surface mixed layer where air-sea exchange processes continuously modify water mass properties. The paper concludes with a brief discussion of the representation of water masses and their evolution in numerical models.

## **1. Introduction**

The existence of water masses has long been recognized in oceanography, and techniques for the analysis of water mass spreading and mixing have been available for decades. Their importance for our understanding of the oceanic circulation has not always been fully appreciated, but recent developments in oceanography and climate research are bringing the study of water masses back into prominence. Oceanographers and meteorologists have come to appreciate the ocean's role in climate variability and climate change and with it the importance of water masses as reservoirs for heat, salt and dissolved gases.

While the surge of papers dealing with the hydrographic structure of the global ocean is welcome, the lack of rigor in the use of the water mass concept can only bring the underlying ideas into disrepute. Ocean modelers and observationalists often use the terms water mass, water type and water body rather indiscriminately to describe property distributions as more or less uniform or following recognizable trends, giving the impression that these terms are more a “manner of speaking” rather than well defined scientific concepts. If the concept of water masses is to make a significant contribution to

1. Flinders Institute for Atmospheric and Marine Sciences, The Flinders University of South Australia, GPO Box 2100, Adelaide SA 5001, Australia *email: matthias.tomczak@flinders.edu.au*

our understanding of the ocean's role in climate processes, it has to evolve from a qualitative description of hydrographic property distributions to a quantitative instrument for the analysis of pathways and budgets.

This paper is an attempt to review and expand the ideas which form the basis of water mass analysis. Its major aim is to strengthen the theory of water masses and open new avenues for quantitative applications. It is divided into the following sections:

- Some notes on the history of water mass analysis
- Physical definitions
- Water mass formation processes
- Quantification of water mass analysis (mathematical definitions)
- Water mass evolution and decay
- Quantitative water mass analysis
- Water type density and Mode Water
- The problem of the oceanic thermocline
- Some remarks on water masses of the surface layer
- Water masses in numerical models
- Outlook

## **2. Some notes on the history of water mass analysis**

The role of water mass analysis in the development of oceanography into a modern science can be described as an evolution through phases. The period before the First World War was dominated by the collection of primary data. Between the wars it was followed by a period of systematic study of ocean basins characterized by ocean-wide expeditions (Tomczak, 1980). The study of water masses formed the major tool for the discovery and description of core aspects of ocean dynamics during this period. Seminal works from this phase of oceanography are still important references. The description of the water masses of the world ocean given by Sverdrup *et al.* (1942) is still unsurpassed and considered an essential reference for today's research (Warren, 1992). Wüst (1936) and Defant (1936) based their description of the circulation of the Atlantic Ocean on a detailed qualitative analysis of the water masses. Their works were translated into English half a century after their original publication (Emery, 1978; 1981), and Wüst's diagram of the water masses and circulation of the Southern Ocean can be found in nearly every introductory textbook of oceanography.

The development of the autonomous recording current meter and of techniques for long term deployment of oceanographic moorings shifted the focus of research toward dynamic phenomena observable through direct current observations which challenged the classical

view of the deep ocean as being dominated by slow, diffuse movement. The discovery of strong time-variable ocean currents, of mesoscale eddies, fronts and internal waves dominated the first decades after the Second World War. Spectral analysis of time series data became a central tool in oceanography, resulting in significant discoveries such as the derivation of the canonical internal wave spectrum (Garrett and Munk, 1975).

Classical water mass analysis was ill equipped to contribute to this phase of oceanography. It continued to play a role in the study of the abyssal circulation where a few dedicated individuals published seminal papers (Reid and Lynn, 1971; Warren, 1983), but it was no longer at the forefront of oceanographic research, at least in what was then called the Western World. Soviet oceanographers, who did not have access to similarly advanced current meters and mooring technology, continued to focus on hydrographic properties. They exploited the possibilities offered by the high vertical resolution of the new CTD technology and produced works of lasting importance on oceanic finestructure and fronts (Fedorov, 1978, 1986). The Soviet school of classical water mass analysis, summarized by Mamayev (1975), faded into the background during this period as well.

The last two decades have witnessed the reappearance of water mass analysis as a respected branch of oceanographic research; one of the invited presentations at the assembly of the International Association for the Physical Sciences of the Oceans (IAPSO) held in 1982 in Halifax was entitled "The re-emergence of water mass analysis." This new phase of oceanography is characterized by our growing awareness of the ocean's role in climate variability and climate change. International research programs such as TOGA (Tropical Ocean, Global Atmosphere), WOCE (World Ocean Circulation Experiment) and CLIVAR (Climate Variability) try to clarify the dynamics of the coupled ocean-atmosphere system. The ocean is the flywheel of the climate machine. Its effect on the operation of this machine requires accurate estimates of turnover rates of all major water masses. As a result, the study of water masses has become an important component of international programs such as WOCE and CLIVAR, which have produced and will continue to produce a large amount of high quality hydrographic data and will give water mass analysis a prominent place again in oceanographic research.

This new age of water mass studies will not be a simple revival of the 1930s. Modern water mass analysis can make use of data and tools not available to the oceanographer of 60 years ago.

On the observational side, the introduction of chemical compounds into the ocean through human activity and the development of techniques for measuring their presence in minute quantities has opened the field of tracer oceanography, and the range of parameters which can be used to trace water masses and analyze mixing between them has grown significantly.

The range of methods useful for water mass analysis has also grown notably over the last few decades. Numerical modeling has become a powerful tool in oceanography and is capable of identifying formation regions and pathways of water masses. Inverse methods can be adapted to extract the maximum amount of information on water masses from the

new hydrographic data set which includes nutrients and tracers. By a combination of the classical hydrographic data set with tracer data it has become possible in principle to derive the age of individual components contained in a mixture of water masses (Karstensen and Tomczak, 1998).

These examples show that water mass analysis is on the threshold of a new era. It will evolve from a tool for the qualitative description of the oceanic circulation into a tool for the quantitative determination of water mass formation rates, age along pathways, and storage capacity for heat, carbon and other components of the climate system. This suggests that the time has come to summarize past and present ideas and to reflect on the potential contribution of water mass analysis to future oceanographic studies.

### **3. Physical definitions**

The role and significance of water masses in the ocean can best be understood by drawing on the analogy of air masses in the atmosphere. The idea of air masses developed empirically in the early nineteenth century and was formulated into a scientific concept in Norway, by the famous Bergen School of Meteorology, during the 1920s (Bergeron, 1959). The concept of air masses is a daily ingredient of weather analysis and forecasting, and most people can relate to it through their experiences of weather changes, the passage of fronts and related strong and sometimes violent winds. The ocean and the atmosphere are both fluids in turbulent motion, and the analogies between the two media are well known and usually appreciated in the context of fluid dynamics (Barkley, 1972; Charney and Flierl, 1981). The existence of submerged eddies at oceanic depths with long life times was documented more than a decade ago when McWilliams (1985) showed numerous examples and described them as abundant but spatially sparse. Nevertheless, oceanographers still tend to associate water masses with slow diffusive movement and seem surprised by the presence of fronts and storms in the oceanic interior.

Air masses are formed in certain regions of the atmosphere. Meteorologists identify four geographic zones (Arctic/Antarctic, polar, tropical and equatorial) and two subdivisions (maritime and continental) in each zone. For example, the Maritime Tropical source region is found between 20° and 30° over the oceans of both hemispheres. The different air masses are carried away from their formation regions by the regional wind systems and form frontal systems where they meet. These fronts are associated with an increase in wind strength and gustiness and contain strong updraughts and downdraughts.

Like air masses, water masses are formed in certain regions of the ocean, advected by the regional current systems and form frontal systems where they meet. Fronts between water masses can also be associated with an increase in current strength and turbulence and with significant vertical motion.

There are, however, two major differences. The first difference arises from the equations of state for the two media. The two most prominent properties of air masses are their temperature and humidity. The density of air is determined nearly exclusively by its temperature; water vapor content plays a very minor role, and its effect on air density is

usually negligible outside the equatorial region. As a consequence, frontal systems between air masses, i.e., strong horizontal gradients in the temperature and water vapor fields, are always associated with strong geostrophic flow along the front. In contrast, the most prominent properties of water masses, temperature and salinity, contribute to the density of water in roughly equal proportions. Frontal systems between water masses are therefore not by necessity accompanied by strong geostrophic flow along the front, and the character of turbulence in oceanic frontal systems can be quite different from the wind gustiness known from atmospheric fronts. If the water mass front is density-compensated (i.e., the temperature and salinity changes across the front balance each other with respect to their effect on density), the front is not associated with strong mean flow. Mixing across the front can then be achieved not only by geostrophic turbulence but to a large degree by the formation of intrusions along isopycnals and by double diffusion.

The second difference relates to time and space scales. Air masses develop and decay on time scales associated with weather, so there is not much purpose in tracing them beyond a time span of more than a few days. Time scales in the ocean are much larger, and water masses are elements of climate rather than weather. This makes it not only feasible but useful to trace their history and the evolution of their properties in much more detail than one would ever contemplate doing with air masses. This requires special data collection programs and special methods of analysis, which are now becoming available.

A few definitions are required as a starting point for a meaningful discussion. We define a *water mass* as a body of water with a common formation history, having its origin in a particular region of the ocean. Just as air masses in the atmosphere, water masses are physical entities with a measurable volume and therefore occupy a finite volume in the ocean. In their formation region they have exclusive occupation of a particular part of the ocean; elsewhere they share the ocean with other water masses with which they mix. The total volume of a water mass is given by the sum of all its elements regardless of their location.

It can of course be argued that defining a water mass as a body of water with a common formation history leaves much room for subjective decisions and does not provide a promising basis for quantitative analysis. But the natural world is a complex organism, and attempts to describe it through a system of definitions and classifications will inevitably leave room for argument. The underlying idea of the definition introduced here is to relate water masses to the dynamics of the ocean. Which physical process is responsible for the formation of a particular water mass is usually much less debatable than attempts to draw boundaries in the ocean on the basis of "typical" property values.

A *source region* or *formation region* of a water mass is that part of the world ocean where a water mass attains its characteristic properties. It represents a finite volume in the ocean, defined as the ocean region where water mass formation is taking place. Note that water mass formation can be seasonal or intermittent in time. The formation region is then given by the sum over all ocean regions where water mass formation occurs at some time during the year.

A *conservative property* is a water mass characteristic or property which has no sources or sinks in the ocean interior. Observed changes in conservative properties can therefore only be the result of mixing with surrounding water.

A *nonconservative property* is a water mass characteristic or property which changes during the life span of the water mass. The reasons for this change can be biochemical, as in the case of nutrients and oxygen, or physico-chemical, as in the case of radionuclides and tracers.

In the usual sense these definitions apply only to water masses below the oceanic mixed layer, which acquire their properties in their source region and are cut off from contact with the atmosphere once they leave their source region. Water masses of the upper ocean which are permanently in contact with the atmosphere cannot be defined in the same way. It is possible and useful to define surface layer water masses if recourse is made to the mathematical concept of water types introduced below.

#### 4. Water mass formation processes

Three physical processes are known to produce water masses; in addition, a fourth category of "formation" can be introduced for convenience.

*Convection* is a formation process where the density of water in the oceanic surface layer is increased by cooling or by evaporation. This leads to a sinking of water to the depth where static stability is again achieved. Typical convection depths of individual sinking events are a few hundreds of meters, but the cumulative effect of many events can reach much deeper. After sinking the newly formed water mass is carried away from its formation region by the current which prevails at depth.

Because convection is a nonhydrostatic process, its horizontal and vertical velocities are of the same order of magnitude and the horizontal scale of the convection cells is determined by the convection depth. In other words, the sinking of water in the water mass formation region occurs initially on quite small space scales (hundreds of meters) and is accompanied by the rising of water from similar depth in the formation region itself. In partly ice-covered regions, mixing continues on these small scales (Kämpf and Backhaus, 1998). In deep ice-free regions, convection is progressively controlled by rotational effects and the convective depth scale expands, with a corresponding increase in the mixing between sinking plumes and the surrounding water (Jones and Marshall, 1993; Send and Marshall, 1995).

In both situations, water mass formation by convection is the result of property modification by contact with the atmosphere and of homogenization by mixing in the convection region, which results in extremely uniform water properties over a large depth range. One consequence of this is that a water mass formed by convection rarely enters life fully saturated with oxygen. Its oxygen concentration is determined by the oxygen content of the sinking water, which is saturated in oxygen, and by the oxygen content of the rising water, which mixes with it.

The small horizontal scales involved in convection make direct observations of the

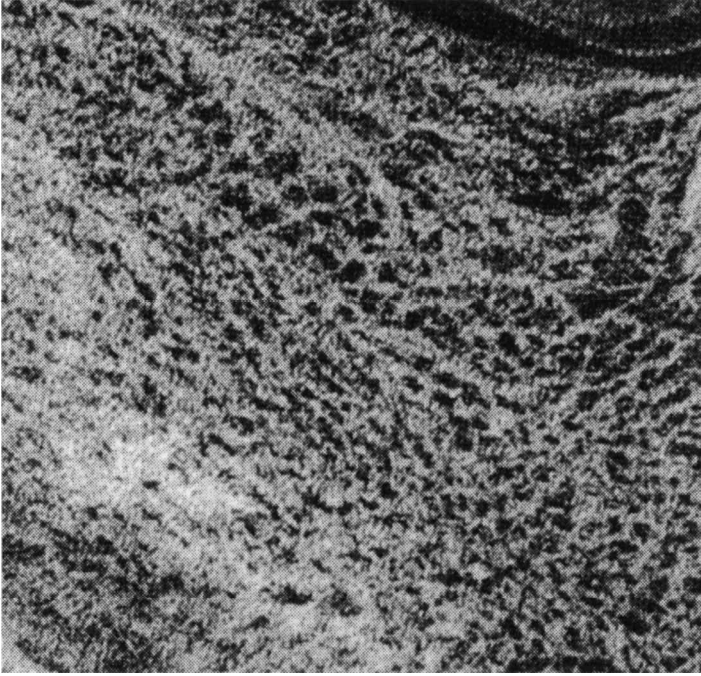


Figure 1. A synthetic aperture radar satellite image of ice cover in the Greenland Sea taken on 12 February 1992. The area shown is approximately  $7.5 \times 7.5$  km in size. Between a lead of open water at the upper right and an area of fresh ice at the lower left is a band of convective cells about 200–400 m in diameter, in good agreement with the expected size of oceanic convection cells. From Carsey and Garwood (1993).

phenomenon difficult in the ocean, but the analogy with the atmosphere can leave no doubt about its ubiquitous occurrence. Convection in the atmosphere is driven by the rising of warm air and made visible by associated cloud formation; the impressive cumulonimbus towers of the tropics give daily demonstrations of the process. Observations of deep homogeneous columns of water indicative of oceanic convection are much rarer; some observations obtained under fortunate circumstances are described by Sankey (1973), Rudels *et al.* (1989), Schott *et al.* (1993) and others. Satellite images of ice cover over the Greenland Sea (Fig. 1) show distinct cellular patterns thought to be the result of convection: The ice accumulates in the surface convergences produced by the convective sinking motion, while the regions of ascending warmer water are kept ice-free.

*Subduction* is a process of water mass formation through a combination of wind action and cooling. A convergence in the wind-driven surface current field (convergence of the Ekman transport resulting from negative wind stress curl, known as Ekman pumping) pushes water down along isopycnal surfaces. This downward motion is extremely slow and during autumn and winter easily overtaken by the downward growth of the mixed layer, which is then convectively mixed as a result of surface cooling. Most of the water carried



downward as a result of Ekman pumping is therefore caught in the mixed layer again during winter, and its properties are then completely uniform. When spring warming sets in, the mixed water is isolated under a thin surface layer of warmer water but continues to move downward on isopycnal surfaces in response to the convergence in the Ekman layer, eventually moving out of reach of next winter's mixing. It is then injected into the ocean interior. The process was first suggested by Stommel (1979) as responsible for the formation of Central Water and has since been verified by Marshall *et al.* (1993), Williams *et al.* (1995) and others. A characteristic of subduction is that freshly subducted water generally is saturated with oxygen.

In practical situations convection and subduction are sometimes found to act together to form a water mass. Observations from the North Pacific Ocean, for example, show that North Pacific Central Water is generally formed by subduction (winter convection followed by downward movement on isopycnals driven by Ekman pumping). In anticyclonic eddies, convection can reach so deep that the resulting homogenization reaches below the Ekman pumping influence, and the homogenized water is trapped in the eddy for months or years. Its movement through the ocean is then controlled by the movement of the eddy at least initially, and it can be found at quite some distance from its formation region as a layer of particularly uniform hydrographic properties often referred to as Mode Water. The major effect of the interplay between subduction and convection is a nonuniform distribution of water types over the range of parameter values encountered in the water. This aspect is discussed in more detail in Section 8.

*Subsurface mixing* is the only formation process which does not rely on air/sea exchange of properties. It is not as common as the other two formation processes. Where it occurs, parts of two or more water masses are brought together and mixed so thoroughly that a substantial depth range of the region where the mixing occurs contains water with well defined properties distinctly different from those of the contributing water masses. This water, which subsequently spreads through the ocean as it is advected by currents, is then considered a newly formed water mass. The most prominent water mass formed largely by subsurface mixing is the Circumpolar Water of the Southern Ocean (Thompson and Edwards, 1981).

On the world ocean scale, the three processes just listed are the only water mass formation processes possible. A fourth formation process, *importation from outside*, has to be considered if the region under consideration is only part of the world ocean. Such a view is often justified in the study of marginal seas. In the European-African Mediterranean Sea, for example, Atlantic Water is a well defined water mass imported through the Straits of Gibraltar. Specification of its formation process is not required for the study of the Mediterranean Sea. Note that even in marginal seas some water masses are formed inside the marginal sea itself. Their individual formation processes have to be specified and can only be one of the three processes defined above.

## 5. Quantification of water mass analysis (mathematical definitions)

The aim of modern water mass analysis is the quantification of formation and mixing rates, total water mass volumes and turnover times. This requires quantitative water mass definitions and analysis methods. This section summarizes mathematical definitions and their relation to water masses in physical space.

Consider a parameter space of  $n$  dimensions defined by  $n$  water properties. In general, temperature and salinity are always included, so the minimum dimension of the parameter space is  $n = 2$ . This is the parameter space of the temperature-salinity ( $T$ - $S$ ) diagram introduced in 1916 by Helland-Hansen. It is, however, often useful to include other parameters, such as oxygen, nutrients, radionuclides and other tracers, and consider water masses in more dimensions. A *water type* is a point in  $n$ -dimensional parameter space; i.e., ie a combination of temperature, salinity, nutrients, oxygen and other tracer values. It is a mathematical construct and does not occupy any volume in physical space. There may be water particles with exactly the properties described by the water type, but the existence of such particles is not necessary for the water type to exist.

If a water type corresponds closely to the properties of a water mass in its source region it is called a *source water type*. Formation conditions in the source region vary over time, and it is again possible that no single water particle has exactly the properties described by the source water type. What this means is that source water types, too, are mathematical constructs and do not exist in physical space. The property combinations which they represent are, however, close to the property combinations found with water particles in a water mass. To be quantitative, the properties of all water particles in the source region are located in a subspace of the  $n$ -dimensional parameter space with the source water type at its center. The diameters of this subspace in the directions of each of the  $n$  axes define the *standard deviations* of the water mass properties.

Observations of hydrographic properties from the source region (i.e., in physical space) deviate from the corresponding source water type (the corresponding point in mathematical space) for a number of reasons. The first to come to mind is instrumental or, in the case of chemical data, analytical accuracy (grouped together as instrumental variability). Its importance varies widely between parameters; but in many instances it is significantly smaller than the variability inherent in the formation process of the water mass (environmental variability), which is the result of different atmospheric conditions from one year to the next. The instrumental variability of a modern CTD for example is typically  $0.001^{\circ}\text{C}$  for temperature and  $0.003$  for salinity, a small fraction of the environmental variability of sea surface temperature and salinity in water mass formation regions. The standard deviations represent the combination of these two types of variability, although in practice they are mostly determined by the environmental variability.

Mathematically a water mass can then be defined by a combination of source water types and associated standard deviations. Some water masses, notably those formed by convection, are characterized by very uniform property distributions in space and require only a

single water type and its  $n$  standard deviations. Examples are Antarctic Bottom Water and North Atlantic Deep Water.

Other water masses, notably those of the permanent thermocline, span a large range of parameter values and are characterized by parameter relationships such as temperature-salinity ( $T$ - $S$ ) curves and corresponding relationships between their other properties (temperature-oxygen, temperature-nutrient, temperature-CFC and so on). Examples are the various Central Waters of the three subtropical and tropical oceans. In  $n$ -dimensional parameter space, relationships between parameters (curves) correspond to an infinite number of source water types. In practical situations it is often possible to approximate these curves by mathematical relationships which require only few points (source water types) to be fully defined. If the relationship is linear (straight lines in all property-property diagrams), two source water types and one set of standard deviations are sufficient to define the water mass. The departure of the observed property values in the source region from the mathematical straight lines is then another reason why observations of water mass properties do not necessarily correspond to parameter combinations defined by source water masses, and the standard deviations include this as an additional effect (together with environmental and instrumental variability).

Figure 2 gives an example of a representation of two water masses in mathematical space, assuming for simplicity a three-parameter representation ( $n = 3$ ). The projection onto the  $T$ - $S$  plane reproduces the classical  $T$ - $S$  diagram.

If the variables used to define the parameter space include tracers with a time dependent input function such as the CFCs or carbon-14, time has to be included as an additional parameter, and the definition of water types and standard deviations has to proceed in  $n + 1$  dimensional parameter space. In practice this has not been attempted yet. The most advanced study of this type (Karstensen and Tomczak, 1998) uses the time independent parameters to define the  $n$ -dimensional parameter space and handles the time dependent parameters one at a time as a one-dimensional time problem. Eventually it will be found that what we consider time independent water type properties today may well turn into time dependent properties when our expanding data base allows us to address changes of water mass properties on climate scales. It will then become necessary to define all water mass properties in  $n + 1$  dimensions.

## 6. Water mass evolution and decay

The previous sections were concerned with water mass formation and mathematical formalisms needed for quantitative description. This section is a review of processes active after water mass formation. The question how these processes can be observed and quantified will be addressed in the next section.

The first process to occur is further homogenization of the water mass through mixing within its interior. This process, properly called *water mass consolidation*, is active even in the absence of any mixing with other water masses; in other words, it occurs also in situations where the water mass has exclusive occupancy of a certain depth range as it

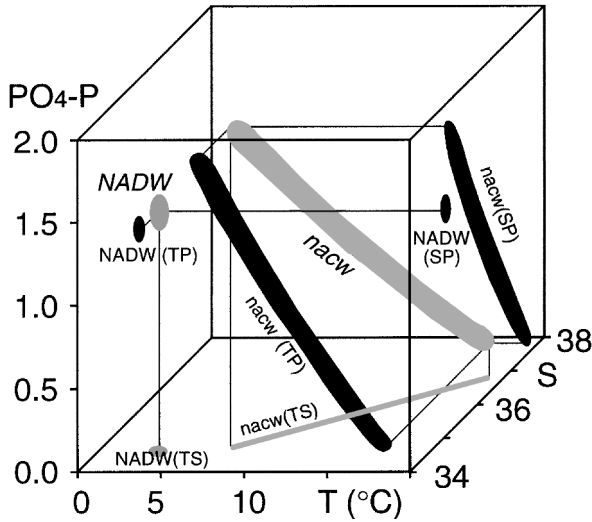


Figure 2. Two representations of water masses in the three-dimensional parameter space spanned by temperature, salinity and phosphate. The figure is loosely based on the properties of North Atlantic Deep Water (NADW) and North Atlantic Central Water (nacw; in departure from normal convention, lower case spelling is used to discriminate between the two water masses in the diagram more clearly). NADW stands as an example of a water mass formed by deep convection and is represented by a subspace centered on a single water type, while nacw is an example of a water mass formed by subduction in the Subtropical Convergence Zone which is represented by a subspace defined by parameter relationships. NADW(xy) and nacw(xy) are the projections of the water types onto the xy-plane. The thickness of a curve or the size of a region indicate the magnitude of the respective standard deviation.

spreads through the ocean. It can take various forms. In the formation region itself it may reduce the degree of interannual variability in the water mass properties by mixing newly formed contributions from one formation event (for example one winter period) with contributions from previous formation events (previous winter periods). As a result, the effect of environmental variability is reduced and the  $n$  standard deviations for the water mass in parameter space are kept small. Where a water mass is characterized by a set of property relationships, water mass consolidation may eliminate layers of large vertical property change and lead to a more uniform distribution of vertical property gradients in space (see the section on water type density below). Departures from smooth relationships may be eliminated, resulting again in a reduction of the associated standard deviations. It has been argued, for example, that double diffusion causes the  $T$ - $S$  curve of Central Water to absorb departures from its long-term mean position in  $T$ - $S$  space by making it converge toward a curve of constant density ratio. This possibility will be addressed in more detail in a later section.

The result of water mass consolidation in physical space is a redistribution of water mass

Table 1. The life cycle of water masses.

formation	convection subduction subsurface mixing importation from outside (on regional scales only)
evolution	consolidation aging mixing
decay	absorption transformation

properties. In mathematical space water mass consolidation reduces the standard deviations without changing the water type definitions of the water mass.

A second process, which is active as soon as a water mass leaves its formation region, is *water mass aging*, which might also be called *water mass modification*. It occurs irrespective of any mixing with other water masses. It has no effect on the conservative properties but changes nonconservative properties in line with their bio-chemical or physico-chemical behavior. As a result, a water mass undergoes change even if it continues to be the only occupant of a well defined region of the ocean. The result of water mass modification is a change of water mass properties in physical space accompanied by a corresponding change of water types in mathematical space.

The next process to occur is *mixing* between water masses. Mixing does not result in a change of water mass properties in physical space but it produces new combinations of water properties not found in the water masses themselves. The mixing process defines a mixing subspace of the  $n$ -dimensional parameter space (the mixing line in the  $T$ - $S$  diagram between two water types, the mixing triangle between three water types and so on). It is always possible in principle to identify the contributing water masses and determine their relative contribution to the new mixture.

The sequence of water mass consolidation, modification (aging) and mixing can be described as the period of water mass *evolution*. Eventually every water mass is mixed with others to such a degree that it can no longer be identified. In some situations the water mass disappears without trace, as for example Mediterranean Water which eventually gets absorbed into North Atlantic Deep Water. This end to a water mass might appropriately be called *water mass absorption*. In other situations, water masses disappear by being transformed into new water masses. Australasian Mediterranean or Banda Sea Water, for example, is formed in the Indonesian seas by transformation of Pacific Central Water. This end to a water mass is therefore called *water mass transformation*. Water mass formation by subsurface mixing is always associated with water mass transformation—the end of one or more water masses defines the beginning of a new one. Water mass absorption and transformation are two possible forms of the final phase in the history of a water mass; they can be classed together as water mass *decay*.

Table 1 gives an overview of the life cycle of water masses and a summary of the

discussion of the previous sections. Most of the terminology is new, has not been applied in previous studies and may appear over-structured; but it is important to realize that the table lists physical processes which result from different dynamics and may therefore require different methods of analysis. Quantitative water mass analysis is only possible if it applies the correct tools, and selecting the correct tool requires an understanding of ocean dynamics.

## 7. Quantitative water mass analysis

This section reviews the observational oceanographer's tools which are useful for the analysis of water masses. The question of water mass identification in numerical models will be addressed in a later section.

Water mass visualization will of course always rely on temperature-salinity and other property-property diagrams. Such graphical methods are invaluable as orientation tools, but their use in quantitative analysis is very limited. Other diagrams, such as plots of temperature or salinity as functions of density against space or time, might be useful to illustrate the effect of water mass mixing. They do not help much when it comes to quantitative analysis. The products most useful for climate analysis will eventually be maps of water mass content (water mass volume per km<sup>3</sup> of ocean), water mass age and water mass formation rate (km<sup>3</sup> of oceanic surface layer per year).

Water mass consolidation redistributes the water mass properties in physical space but does not affect the source water types. Because the essence of *T-S* diagrams is to display a subset of the *n*-dimensional water type distribution in parameter space, it cannot show the effect of water mass consolidation. Consider the example of a water mass formed by winter convection and then advected slowly into the ocean interior. Assume that its properties show no interannual variability. Assume that a volume *V* is formed every winter and mixes with the volumes of the previous *k* years along the advection path. This will not change the *T-S* properties of the water mass; but it will determine the age of this water when it eventually comes into contact with and mixes with other water masses. What a *T-S* diagram shows as the starting point for the mixing process are the *T-S* properties of the entire volume (*k* + 1)*V*. This information can be used to study the mixing path of this water mass when it begins to mix with other water masses; but to determine the role of this water mass in air/sea exchange it is also essential to determine the individual annual contributions contained in this volume.

The question can be addressed through the use of tracers which have a time-variable input function (Roether *et al.*, 1993; Roether, 1994). Water mass consolidation then becomes visible in the tracer data because the tracer concentration within the water mass is changed by the addition of newly formed source water at different points in time, even though temperature and salinity remain the same. This leads to the concept of an "age profile" for a water mass, which quantifies the contributions of all formation events to the total water mass volume and gives information on the most common age and on the age span of given percentiles of a water mass (Putzka, 1995). Just as an age profile of a

population is a good indicator of a nation's place in the global economy, the age profile of a water mass is an indicator of its place in the global climate system.

The analysis of water mass modification (aging) poses different problems. The fact that nutrients are brought back into solution and utilize oxygen in the process can be used quantitatively if a few assumptions are made. The starting point is the observation that the phosphate and nitrate content increases with water mass age in a fixed ratio to the decrease of oxygen content, known as the Redfield ratio (Redfield *et al.*, 1963). If an estimate for the oxygen consumption rate per unit time (year, decade, century) can be found, the observed changes of nutrients and oxygen can be converted directly into water mass age. Optimum Multiparameter (OMP) Analysis (Tomczak, 1981a; Mackas *et al.*, 1987; Tomczak and Large, 1989) can achieve the same if the Redfield ratio and water mass age are included in the water mass matrix (Poole and Tomczak, 1999; Karstensen, 1998).

Without knowledge of the oxygen consumption rate both methods (direct and OMP) produce a relative age, a somewhat unsatisfactory quantity from the theoretical point of view but still useful in many situations. A core assumption is that the Redfield ratio is accurately known and constant over the region under consideration. There is enough evidence to suggest that it varies between ocean regions and with depth (Takahashi *et al.*, 1985; You and Tomczak, 1993). It is possible to adapt the existing methods to an ocean with a variable Redfield ratio, but it is unclear whether enough is known about the mechanisms behind the observed variability to justify this step.

A crucial assumption for all age determinations is that the water mass is not mixing with any other water. If the water parcel under scrutiny is a mixture of water masses, the determined age is an artificial and rather meaningless number representing the "mixed age" in the water parcel. It is of course of interest to trace the evolution of individual water masses beyond the point of mixing. Numerical models do not have a difficulty here; the introduction of water mass tracers in combination with a Lagrangian approach allows age determination for individual water masses. Data analysts are faced with the inverse problem of finding the age of individual components in an observed mixture from the knowledge of the "mixed age." It can be shown that linear inversions, such as OMP analysis with the inclusion of the linear Redfield ratio relationship between nutrients and oxygen, do not produce a uniquely determined solution of this problem. Tracers such as CFC-11 and CFC-12, which have an input function with a strongly nonlinear time dependence, can be used to overcome this problem. Karstensen and Tomczak (1998) demonstrate that it is possible to determine the individual ages of Central Water and Subantarctic Mode Water in the mixing region between these two water masses through a combination of linear OMP analysis and an inversion of CFC-11 and CFC-12 observations.

The analysis of water mass mixing, the third and last stage of water mass evolution, is relatively straightforward in comparison. The classical mixing diagram in the *T-S* plane produced many valuable results. Its quantitative applications are limited to situations with not more than three water types. This restriction was overcome by the introduction of OMP analysis, which expanded the classical use of temperature and salinity as mixing indicators

and included all parameters available in the  $n$ -dimensional parameter space. This also opened the possibility to distinguish between isopycnal and diapycnal mixing (Tomczak, 1981b), although the error margins are still fairly large (Klein and Tomczak, 1994).

In its initial applications the OMP method ignored the nonconservative character of many parameters and was therefore restricted to applications in small oceanic regions, for example across frontal zones. World-wide application for the tracing of water masses from formation to decay became possible through the use of preformed nutrients (Broecker and Takahashi, 1985; You and Tomczak, 1993) in the OMP matrix. The inclusion of the Redfield ratios in the OMP matrix is equivalent to the use of preformed nutrients but has the added advantage of allowing the determination of water mass age. It is, therefore, now preferred over the use of preformed nutrients (Karstensen, 1998).

## 8. Water type density and Mode Water

This section turns to the problem of the distribution of water masses in physical space. Water types and standard deviations define the locations of water masses in parameter space; they do not by themselves carry information on the distribution of water masses in the ocean. A  $T$ - $S$  diagram for example does not provide depth information unless special annotation of the  $T$ - $S$  curve is included, and even then the information cannot be very detailed. There are, however, instances where continuous depth information along the  $T$ - $S$  curve is of interest, for example if there is reason to believe that the water mass contains a particular type of finestructure or that a certain  $T$ - $S$  “mode” contributes a particularly large volume.

It is possible to distinguish between layers of high vertical uniformity and layers of rapid vertical property change by introducing the concept of *water type density*. Although the property-property relationships define an infinite number of water types, water type density is a well defined property of the water mass. It is defined as the normalized (nondimensional) total property gradient along the vertical in physical space and expresses the degree of homogeneity in the water mass as a function of depth.

Mathematically, and expressed in finite difference form suitable for applications to hydrographic data, water type density  $\delta_{wt}$  is defined as follows. Assume that a hydrographic station extends over the depth range  $r$  and the observational increment is  $\Delta z = r/m$ , where  $m$  is a reasonably large integer. The accumulated change  $L_j$  of property  $p_j$  over the range  $r$ , expressed in finite difference form, is then given by

$$L_j = \sum_{i=1}^m (|p_j(z_i) - p_j(z_{i-1})|)$$

where  $z_i$  ( $i = 0, m$ ) represents the depth increments over the range  $r$ . The contribution of property  $p_j$  to  $\delta_{wt}$  is then

$$\delta_j = \frac{|p_j(z_i) - p_j(z_{i-1})|}{L_j}$$



and the water type density  $\delta_{wt}$  is found by adding the contributions from all properties:

$$\delta_{wt} = \sum_{j=1}^n \delta_j.$$

$L_j$  can be interpreted as the length of the water mass definition curve in parameter space with respect to parameter  $j$  (as seen, for example, in Fig. 2), expressed in dimensional units. Water type density is then the nondimensional contribution of the ocean layer between  $z_{i-1}$  and  $z_i$  to the total length  $L_j$ ; it establishes the link between physical space and mathematical space.

A similar concept was introduced by Moiseyev and Chertushkin (1995) for quantitative measurements of finestructure in  $T$ - $S$  space. These authors also discuss possible higher order tensor formulations.

Figure 3 gives two examples of water type density. Both use the two basic conservative water mass properties (temperature and salinity) only. There is no reason in principle why the calculation cannot include other variables if they are available at sufficient vertical resolution, but the focus in the present context is not on this extension of the concept. The figure is presented to demonstrate the different internal structure of water masses in the permanent thermocline. The data from the western Pacific Ocean show a well developed step structure of alternating layers of low and high water type density. This is an indication that double diffusion is active in Pacific Equatorial Water. In contrast, the data from the southeastern Indian Ocean show an extensive layer of very low water type density indicative of the presence of Subantarctic Mode Water.

This leads to the definition of *Mode Waters* as layers of minimum water type density of substantial vertical extent. Note that this definition links Mode Water with physical space; Mode Water is thus not a mathematical construct but a physical entity with a measurable volume.

The world ocean contains many examples of layers of minimum water type density. Many of these are known as Mode Waters, particularly if they can be traced to the surface and linked to a water mass formation process. Subtropical (Masuzawa, 1969) and Subpolar (McCartney, 1977) Mode Waters are formed by oceanic heat loss during winter. There is also a Tropical Mode Water which has not received much attention and still lacks its own name. It is formed by evaporative water loss in the central South Pacific Ocean and spreads with the South Equatorial Current westward at 200 m depth.

Mode Waters are often described as water masses in their own right. This is clearly justified if we recall our definition of a water mass (a body of water with a common formation history, having its origin in a particular region of the ocean), but it is necessary to recall that the departure point for the definition of Mode Waters was the observation of minima in water type distributions, without consideration of particular water mass formation processes. It is possible that the formation process which creates the water type minimum is no different from the formation process which acts on either side of the

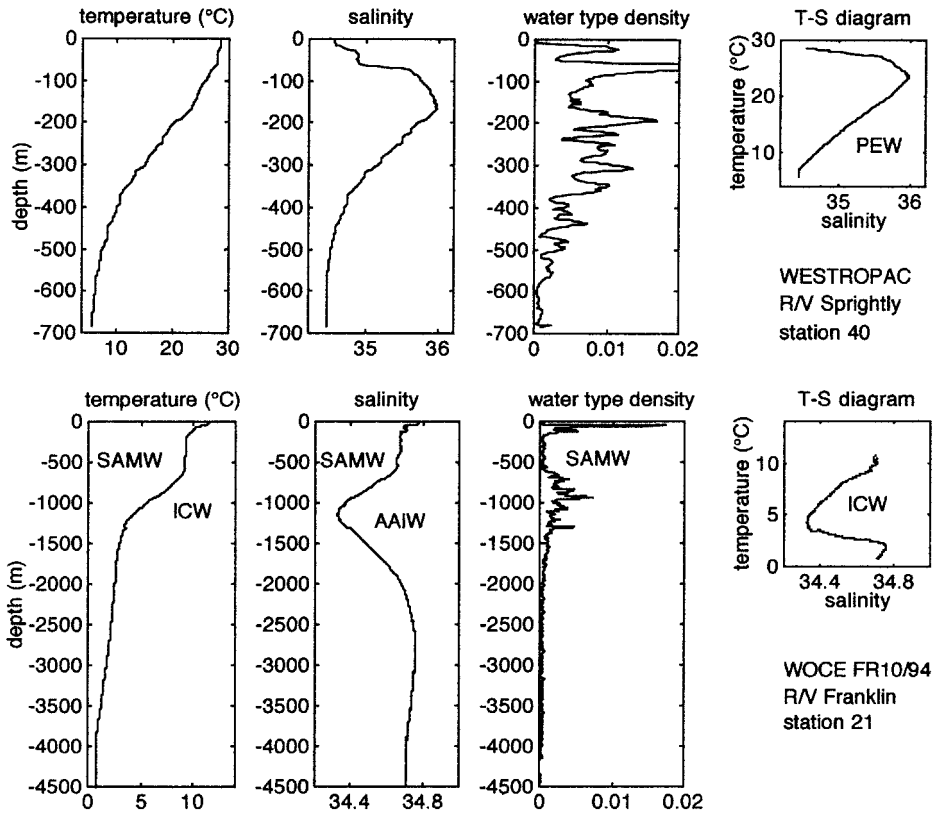


Figure 3. Temperature, salinity and water type density as functions of depth. (top) WESTROPAC station 40 at 11°32'S, 169°52'E, (bottom) WOCE FR10/94 station 21 at 42°S, 120°42'E. CTD data are sampled every 2 meters, water type density is calculated from temperature and salinity only with the length of the  $T$ - $S$  curve measured from the surface to maximum sampling depth (680 m for the WESTROPAC data, 4600 m for the WOCE data). A running average over 20 m was applied to water type density. Things to note are: the much larger water type density in the tropical/subtropical thermocline (in this case Pacific Equatorial Water (PEW) at 11S) compared to the subpolar region (Indian Central Water (ICW) at 42S); the very low data type density in the deep water masses below the level of Antarctic Intermediate Water (AAIW); and the very low water type density in the Subantarctic Mode Water (SAMW) which does not show up in the  $T$ - $S$  diagram, where SAMW is included in ICW.

minimum. Take, for example, Subantarctic Mode Water (McCartney, 1977). It is formed in the Subantarctic Zone just north of the Subantarctic Front and follows the general subduction path of Central Water into the ocean interior, with which it shares a common formation history. It can therefore justifiably be described as the low temperature component of Central Water. In the extreme eastern Pacific Ocean it attains a temperature so low that it becomes the upper temperature component of Antarctic Intermediate Water.

In both cases it can be traced in the ocean as a layer of minimum water type density within an established water mass.

Most subtropical and subpolar Mode Waters are embedded in the water masses of the permanent thermocline and can be considered part of them. The fact that they are associated with a water type minimum makes them particularly easy to trace. It is, therefore, often useful to name them separately and regard them as water masses in their own right.

## 9. The problem of the oceanic thermocline

This section addresses the process of water mass consolidation in the oceanic or permanent thermocline. The water masses in this depth range, known as Central Water, are formed by subduction in the Subtropical Convergence. This is a wide region of the subtropical and temperate climate zones characterized by negative wind stress curl and thus Ekman layer convergence, which produces downward transport of water from the surface mixed layer (Ekman pumping).

As a result of its formation process, Central Water is characterized by a  $T$ - $S$  curve which reflects the meridional variation of temperature and salinity in the surface mixed layer of the Subtropical Convergence (Sprintall and Tomczak, 1993). The degree of water mass consolidation in the oceanic thermocline becomes obvious when a typical Central Water  $T$ - $S$  curve is compared with a typical  $T$ - $S$  diagram produced from a meridional surface profile: Although it is possible to describe both curves as nearly linear in  $T$ - $S$  space and define them through two water types, the environmental variability in the surface layer is much larger (Fig. 4). This is evidently the result of atmospheric variability, which determines the  $T$ - $S$  combinations of the mixed layer and imposes its short space and time scales on it. The mixed layer is, however, unable to transmit these scales of variability into the Central Water; water mass consolidation eliminates them from the water mass.

The question then arises which process is responsible for the observed water mass consolidation? One widely accepted answer goes back to Ingham's (1966) argument that the  $T$ - $S$  curve of Central Water is better described by a constant density ratio  $R_\rho$  than by a linear  $T$ - $S$  relationship. The constant density ratio  $R_\rho$  is defined as  $R_\rho = (\alpha\Delta T)/(\beta\Delta S)$ , where  $\alpha$  is the coefficient of thermal expansion ( $\alpha = 1/\rho \partial\rho/\partial T$ ) and  $\beta$  the coefficient of haline contraction ( $\beta = 1/\rho \partial\rho/\partial S$ ) and  $\Delta S$  is the salinity change along the  $T$ - $S$  curve for a given temperature change  $\Delta T$ . This idea formed the basis of Schmitt's theoretical reasoning (Schmitt, 1981, 1990) that the consolidation of the scattered  $T$ - $S$  relationship of the surface layer into a constant density ratio curve is the result of double diffusion. Following the same argument, Figueroa (1996) presents a world survey of  $T$ - $S$  curves from the oceanic thermocline and concludes that "most of the ocean's central water  $T$ - $S$  curves are better described by a constant density ratio  $T$ - $S$  curve than by a straight line."

The idea is appealing, and until we find a better or alternative explanation it is reasonable to accept it as a realistic hypothesis. Schmitt's theoretical argument seems sound; so the

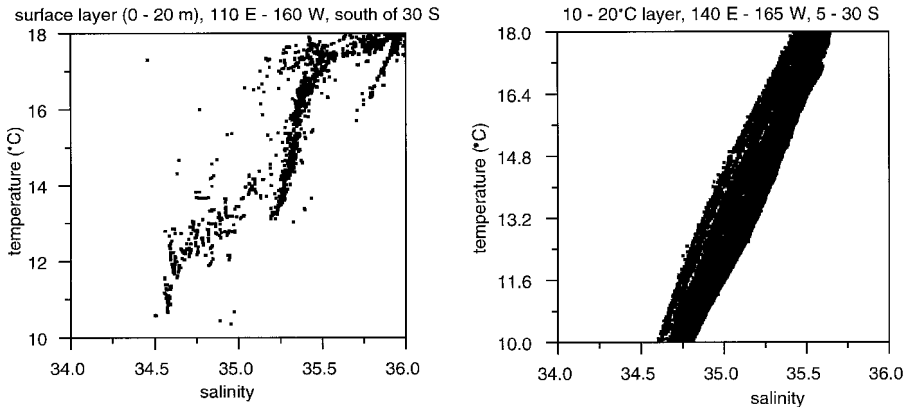


Figure 4. A comparison of space variability in the surface  $T$ - $S$  properties of the Subtropical Convergence (left) and in the  $T$ - $S$  properties of Central Water (right). The surface observations are from the eastern South Indian and western South Pacific Ocean and are based on cruises from various vessels (*Eltanin*, *Diamantina*, *Gascoyne*, *Sprightly* and *Franklin*) over the period 1960–1994. Most of these data were collected during summer, and the data density for some seasons is still poor. The thermocline observations of Central Water are from the western South Pacific Ocean and were collected by R/V *Franklin*, R/V *Malcolm Baldrige* and R/V *Knorr* during the years 1988–1994. At least three distinct bands of well-defined  $T$ - $S$  relationships can be seen, reflecting slight differences of Central Water properties in the Tasman Sea, the Coral Sea and the region east of New Zealand.

test of the idea lies in verification of Ingham's and Figueroa's argument. This requires some reflection on the physical processes at work during water mass consolidation.

What can the  $T$ - $S$  relationship tell us about the type of mixing that occurs in the water column? Theoretical considerations suggest that double diffusion produces  $T$ - $S$  curves characterized by a constant density ratio  $R_\rho$ . Turbulent mixing, on the other hand, affects temperature and salinity in the same way and therefore produces straight lines between the  $T$ - $S$  points of the water bodies involved in the mixing, *provided the supply of water from the water masses involved is infinite*. If a water mass is present only in finite volume the straight mixing line cannot be maintained and changes into a curve. An obvious example is Antarctic Intermediate Water (AAIW). Figure 5 demonstrates how the two straight mixing line connections between AAIW and Central Water and between AAIW and Deep Water change into a  $T$ - $S$  curve which above the level of the AAIW salinity minimum is concave downward ( $\partial^2 T / \partial S^2 < 0$ ).

The problem is that constant density ratio  $T$ - $S$  curves are also concave downward and that considerable skill is required to distinguish between concavity produced by a constant density ratio and concavity produced by the erosion of the AAIW salinity minimum. Figueroa defines Central Water as "the waters between the near-surface salinity maximum and the intermediate waters salinity minimum." He then proceeds to find the best constant  $R_\rho$  fit for the  $T$ - $S$  curve between these two salinity extrema. This procedure cannot produce

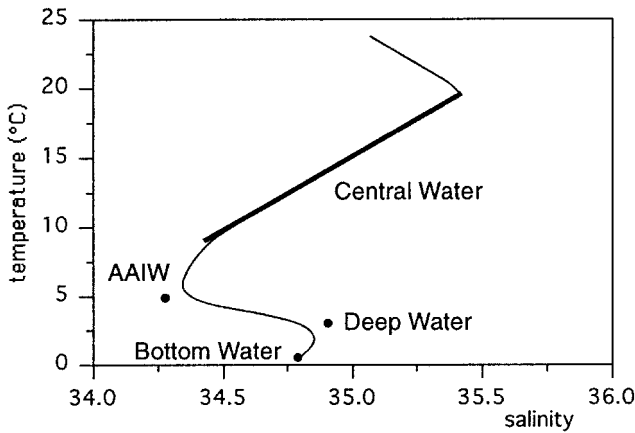


Figure 5. A schematic  $T$ - $S$  diagram for the world ocean. Antarctic Intermediate Water (AAIW), Deep Water and Bottom Water are produced by convection and form homogeneous water masses represented by  $T$ - $S$  points. Central Water is produced by subduction and forms a layered water mass represented by a  $T$ - $S$  curve. Mixing between these water masses and surface layer influence produce the typically observed  $T$ - $S$  diagram shown as a thin line.

reliable information on the shape of the  $T$ - $S$  curve in the Central Water because the result is strongly influenced by the concavity produced by the AAIW salinity minimum. The constant density ratio fit will then always be superior to the straight line fit, but not necessarily because of the shape of the  $T$ - $S$  relationship in the Central Water.

An argument in support of Figueroa's approach could be that the double diffusive regime, in which temperature and salinity both decrease with depth, extends to the depth of the AAIW salinity minimum. Double diffusion may therefore not only be instrumental in shaping the  $T$ - $S$  curve of Central Water but also control the transition from Central Water into AAIW. There is observational evidence that this may not be so.

Figure 6 shows a  $T$ - $S$  diagram from the central South Pacific Ocean, taken from the data supplied with OceanAtlas (Osborne *et al.*, 1991; Osborne and Swift, 1996). A best  $R_\rho$  fit over the range from the salinity minimum to the maximum would certainly be superior to a straight line fit over the same range. However, as suggested by the lines in the diagram, an alternative interpretation could be a linear  $T$ - $S$  relationship in the Central Water and turbulent mixing along a straight line between AAIW and the lower end of the Central Water  $T$ - $S$  range. It can of course be argued that the lack of data in the critical range 8–12°C does not support the straight line interpretation. But the change of slope near 10°C also shows up in the highly smoothed Levitus data. Figueroa's error ratio for the location is 0.43, implying that the constant density ratio fit is superior to the straight line fit by more than a factor of 2. Similar examples based on modern CTD data can be found for other ocean regions. Figure 7 gives a recently obtained example from the Indian Ocean at a location where Figueroa gives an error ratio of 0.37 (preference of the constant density ratio fit over the straight line fit of over 2.7).

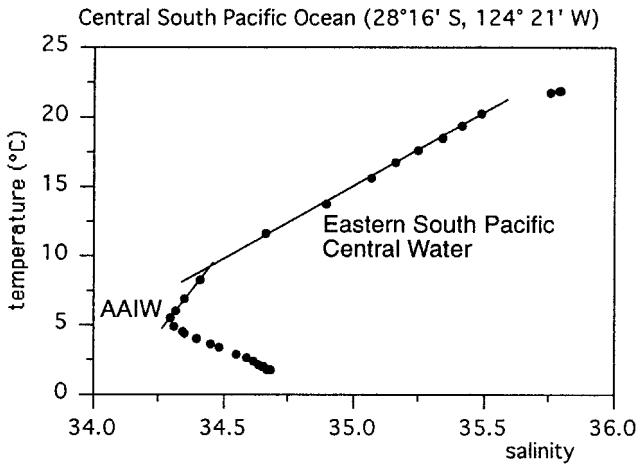


Figure 6. *T-S* diagram from the central South Pacific Ocean. Dots show bottle data. The two lines show a possible interpretation of the *T-S* curve.

The question what determines the shape of the *T-S* curve of Central Water has not received a satisfactory answer yet. Further study of climatological mean *T-S* curves may reveal more information. Tomczak and Gu (1987) showed that in a relatively small part of the equatorial western Pacific Ocean, regions where the constant  $R_p$  model is superior to the straight line (constant  $\Delta T/\Delta S$ ) model are found side by side with regions where the opposite is true. They demonstrated that on occasions a layer with constant  $R_p$  is found above or below a layer of constant  $\Delta T/\Delta S$ . Most of their findings could be explained in

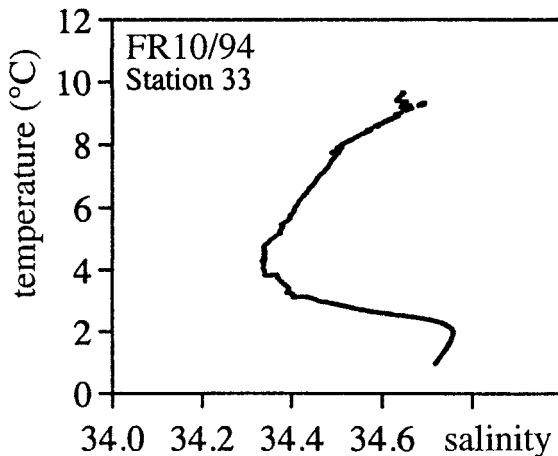


Figure 7. *T-S* diagram from the Great Australian Bight, obtained during R/V *Franklin* cruise FR10/94 (station 33 at 48S, 122E). Notice the clear break in the slope of the *T-S* curve near 8°C. The nearly straight line in the range 8–9°C is Indian Central Water (ICW), the nearly straight line in the range 5–8°C could be the result of mixing between AAIW and the lower end of ICW.

terms of ocean physics; for example, the highly turbulent Equatorial Undercurrent is associated with constant  $\Delta T/\Delta S$  rather than constant  $R_p$ .

## 10. Some remarks on water masses of the surface layer

The concept of water mass evolution can only make sense if temperature and salinity can be defined as true conservative properties, in other words after a water mass is no longer in contact with the atmosphere. A discussion of water masses thus usually excludes the surface layer of the ocean and considers it only in the context of water mass formation.

On the other hand, there is a growing need for some conceptual framework for the classification of water masses in the oceanic surface layer. The amount of oceanic surface data will increase rapidly with the expansion of satellite oceanography. Remote sensing of sea surface temperature is already routine. The launch of SeaWiFS added near-global coverage of phytoplankton concentration and other bio-optical mixed layer properties to the data base (<http://seawifs.gsfc.nasa.gov/SEAWIFS.html>). Even sea surface salinity, long believed to be beyond the reach of remote sensing technology, is likely to be available from satellites soon (Lagerloef *et al.*, 1995). The explosion of data requires some mathematical framework of analysis which can relate the observations to physical processes.

The concept of water types is a good starting point for a definition of water masses in the oceanic surface layer and in shelf regions where the water is in constant contact with the atmosphere. For such a situation a water type is by necessity defined as a point in  $n + 1$  dimensional space, where the additional dimension is time. An *upper ocean water mass* is then defined by the space-time properties of a water type and associated time-dependent standard deviations. In practice, only parameters with a time-independent climatological mean are considered and the time dependence of the water type is conveniently expressed through a list of 12 mean monthly property sets which describe the annual cycle of water mass properties and 12 associated sets of standard deviations which measure the interannual variability.

Most applications restrict the concept to temperature and salinity and define surface water masses through  $T$ - $S$  time diagrams; Figure 8 shows some good examples. But there is no reason why other properties such as chlorophyll- $a$  as an indicator of the productivity of a water mass cannot be included and the concept expanded to  $n + 1$  dimensions.

In the same way as the parameters which define the  $n$ -dimensional parameter space for subsurface water masses are defined through their relevance for the evolution of water masses, parameters for the definition of the  $n + 1$  dimensional parameter space for surface water masses have to be selected in accordance with their relevance for the time history of the surface water masses. Oxygen, for example, is likely to be near or above saturation level at all times in the surface layer and is therefore not a very descriptive parameter for surface water masses. Of much more relevance will be parameters such as the Ekman pumping velocity, the evaporation/precipitation balance, the heat flux, wind speed and

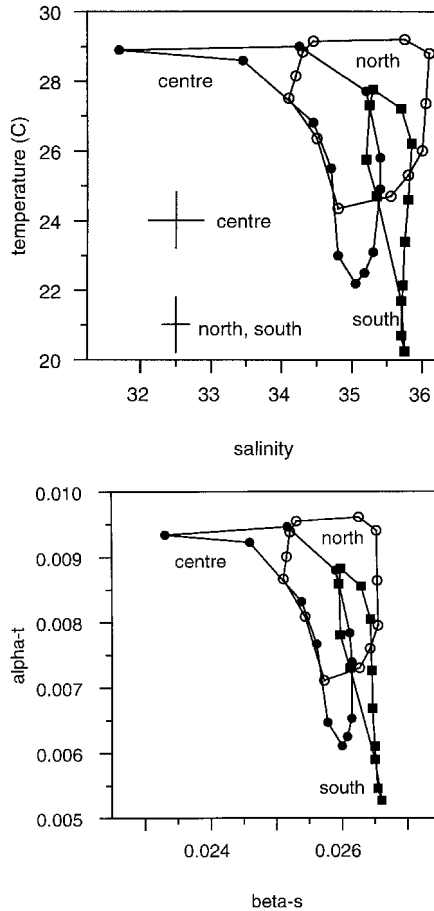


Figure 8. Water masses of the Great Barrier Reef lagoon. The three diagrams show the variation of temperature and salinity during one year for the northern (circles), central (dots) and southern region (squares), approximately defined as north: 10–15S, center: 15–20S, south: 20–25S. Data points are given for each month. (a)  $T$ - $S$  time diagrams. The coldest water is encountered during July; other months are indicated by markers along the curves. The shift of the temperature range indicates the location of the water mass relative to the equator. Salinity minima during summer months (January–March) reflect the effect of monsoonal river run-off. Note that lack of data did not allow calculation of monthly standard deviations; mean annual standard deviations are given instead. Adapted from Pickard *et al.* (1977). (b)  $\alpha T$ - $\beta S$  time diagrams. The salinity contribution to seasonal density changes is always smaller than the temperature contribution, but only marginally so in the monsoon affected region.

others. Their mean annual cycle and interannual variability will have to be included in meaningful definitions of surface water masses.

Because upper ocean water masses undergo permanent renewal, their life cycle is fully described by the space-time properties given through its water types. The concept of



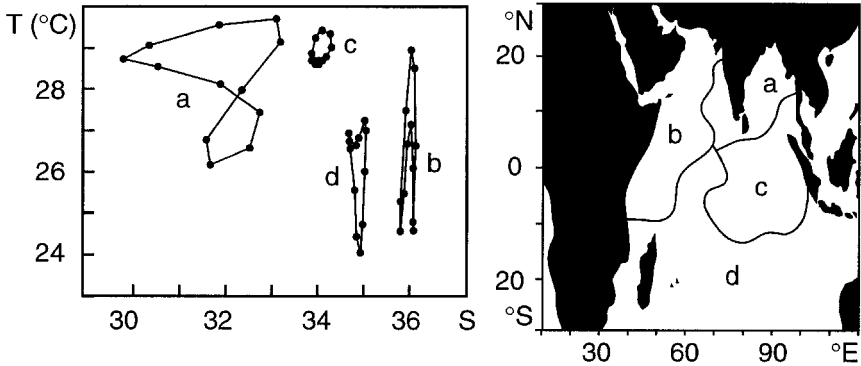


Figure 9. Surface water mass regions of the Indian Ocean based on water type time diagrams for the surface layer. (a) Bay of Bengal Surface Water, (b) Arabian Sea Surface Water, (c) Equatorial Indian Ocean Surface Water, (d) South Indian Ocean Surface Water. The regions are defined on temperature, salinity, chlorophyll, Ekman pumping velocity, net freshwater flux, zonal wind stress component, meridional wind stress component, and wind stress magnitude. The diagram on the left gives typical  $T$ - $S$ -time diagrams for the four regions. Dots indicate months. Note the prominence of the semi-annual cycle in the salinity in the Bay of Bengal (a) and in the temperature of the Arabian Sea (b). Based on Carvalho (1998).

formation, evolution and decay and the associated processes listed in Table 1 cannot apply to them. The concept of water mass density, which quantifies variations of water mass properties in the vertical space coordinate, is replaced by a concept of quantification of changes in time. Changes of temperature and salinity are of particular interest in this context because they determine sea surface density and therefore the stability of the surface layer. To measure their respective contributions to density change it is necessary to convert both into nondimensional properties. Temperature  $T$  is thus replaced by nondimensional or densometric temperature  $\alpha T$  and salinity  $S$  by nondimensional or densometric salinity  $\beta S$ . As Figure 8 shows, the corresponding  $\alpha T$ - $\beta S$  diagram allows direct and easy evaluation of the relative roles of  $T$  and  $S$  with respect to density changes over the seasons.

Preliminary work in the Indian Ocean indicates that the combination of these new surface water mass parameters with the classical  $T$ - $S$  time diagram leads to a natural regional division into clearly distinctive surface water masses (Fig. 9). In the present context these considerations can only have the form of preliminary remarks, but there is little doubt that the concept of water types in  $n + 1$  dimensional parameter space can produce objective and meaningful water mass definitions for the oceanic surface layer.

## 11. Water masses in numerical models

Historically, numerical models of the oceanic circulation began by addressing problems related with the wind driven circulation. They could therefore be quite successful in reproducing known features of the upper ocean circulation without much regard to details of the temperature and salinity fields. Over the last two decades some well established

ocean general circulation models were coupled to atmospheric models. While the inability of some models to reproduce water mass formation processes correctly was not of great concern in the uncoupled mode, their use in the coupled mode can result in unrealistic feedback between the two media. Yet there have been situations where a water mass such as Antarctic Intermediate Water was completely absent from the oceanic circulation of a coupled model (England *et al.*, 1993).

The situation has much improved during the last decade; most numerical models of today are analyzed with respect to the occurrence and depth of convection and with estimates of subduction rates. Many models restore the interior temperature and density fields of the model to observed mean *T-S* fields for reasons of computational efficiency. Because global mean temperature and salinity data sets are carefully screened for instabilities in the density field, restoring model fields to such observations will inevitably underestimate convection, one of the most important water mass formation processes. There is, therefore, not much gained by the introduction of advanced water mass tracers into numerical models if the temperature and salinity fields are not allowed to evolve freely. Recent modeling efforts have concentrated on computationally efficient yet physically correct representation of convection and subduction (England and Maier-Reimer, 1999).

Representation of mixing processes in numerical models has also made great advances, based on theoretical analysis which estimates the contributions from different processes to isopycnal and diapycnal mixing. The arguments are usually developed in terms of the velocity field but can be seen in the context of water mass evolution as well. Most parameterizations are based on the concept of eddy mixing as the major mixing agent (Gent *et al.*, 1995). Internal waves require a different parameterization (Tandon and Garrett, 1996). The relative importance of the two processes in water mass evolution and decay is not clear at present.

A great advantage of numerical models in the area of water mass analysis is their ability to trace the path of water masses with relative ease by introducing artificial conservative or non-conservative tracers. Modelers have full control over the placement and aging properties of their tracers and can adapt them to any particular problem. This allows them to tag water masses and follow them through the world ocean. Questions such as the relative role of cross-frontal mixing versus surface convection in the Southern Ocean can be addressed by placing different tracers on either side of the front and following their development (Ribbe and Tomczak, 1997). It is even possible to carry an age parameter through the model which measures the time elapsed since a water parcel had last contact with the atmosphere (Haidvogel and Bryan, 1992; England, 1995) or since it left its formation region.

These rapid advances with water mass representation in ocean circulation models have lead to a situation where it is becoming more and more difficult to verify model results against observations. Observationalists do not have the means to match all model findings

with numbers based on field data. One aim of the theoretical ideas presented here is to establish a framework which allows observationalists to derive quantities from their data which can be used to compare model predictions with information derived from hydrographic observations.

## 12. Outlook

The dynamics of ocean and atmosphere are governed by a number of conservation equations: conservation of energy, conservation of mass, conservation of heat, conservation of salt. Different approaches to a complete description of the present and future state of the two media place different emphasis on these conservation principles.

The geostrophic description of the oceanic circulation has been extremely successful (Godfrey, 1989) and is the yardstick for all inverse modeling of the oceanic circulation. Geostrophy assumes a frictionless ocean; it emphasizes conservation of energy and mass at the neglect of conservation of heat and salt. Water mass analysis is based on the conservation of heat, salt and mass but does not consider conservation of energy. Both methods produce valid, mostly complementary and sometimes overlapping results. To mention but one example, the pathways of Central Water in the Atlantic Ocean evidenced through the water mass analysis of Poole and Tomczak (1999) could just as well have been derived from geostrophy. The full dynamics of the ocean circulation are not resolved until the principles of geostrophy are combined with the processes of water mass formation, evolution and decay in a single theoretical framework.

How this will be done is not clear at present, but some possibilities are starting to appear. Potential vorticity is a quantity derived from principles of conservation of energy, and it has been shown from principles of geophysical fluid dynamics that under conditions of geostrophy it, too, is a conserved quantity of the flow field. From the point of view of water mass analysis potential vorticity can be considered a conservative property and used as a water mass tracer. This indicates that the parameter space of water mass analysis can be widened to include properties derived from dynamical principles.

*Acknowledgments.* The ideas described in this paper grew over many years. Many colleagues contributed to their formation through discussion and criticism. Among the more memorable of these occasions is a discussion with Jürgen Willebrand, who suggested the use of potential vorticity as a conservative tracer.

This paper is an expanded version of a manuscript which served as a working paper for a workshop "Water mass analysis as a tool for climate research" at the Joint IAMAS/IAPSO assembly in Melbourne during July, 1977. Discussions during the workshop, particularly with Alfred Putzka, Matthew England and Mark Warner contributed greatly to the final version of the paper.

Thanks go to Toshio Suga for pointing out the role of convection in anticyclonic eddies for the formation of Mode Water and to Thomas Haine for clarifying aspects of oceanic convection.

Comments and suggestions from an anonymous reviewer were highly appreciated and contributed to the final formulation of some aspects of the paper.

## REFERENCES

- Barkley, R. A. 1972. Johnston's Atoll's wake. *J. Mar. Res.*, *30*, 201–216.
- Bergeron, T. 1959. Weather forecasting, *in* *The Atmosphere and the Sea in Motion*, B. Bolin, ed., Rockefeller Institute Press and Oxford University Press, NY, 440–474.
- Broecker, W. and T. Takahashi. 1985. Sources and flow patterns of deep-ocean waters deduced from potential temperature, salinity, and initial phosphate concentration. *J. Geophys. Res.*, *90*, 6925–6939.
- Carsey, F. D. and R. W. Garwood. 1993. Identification of modeled ocean plumes in Greenland gyre ERS-1 SAR data. *Geophys. Res. Lett.*, *20*, 2207–2210.
- Carvalho, O. 1998. A classification of the water masses in the oceanic surface mixed layer of the tropical Indian Ocean based on physical and biological indicators, Ph.D. thesis, Flinders University of South Australia, Adelaide, 242 pp.
- Charney, J. G. and G. R. Flierl. 1981. Oceanic analogues of large-scale atmospheric motions, *in* *Evolution in Physical Oceanography*, MIT Press, Cambridge, MA, B. A. Warren and C. Wunsch, eds. 504–548.
- Defant, A. 1936. Schichtung und Zirkulation des Atlantischen Ozeans: Die Troposphäre. *Wissenschaftliche Ergebnisse der Deutschen Atlantischen Expedition "Meteor" 1925–1927*, vol. VI part 1, 289–411.
- Emery, W. J. 1978. *The Stratosphere of the Atlantic Ocean* by Georg Wüst, U.S. Department of Commerce, Springfield, VA.
- 1981. *Stratification and Circulation of the Atlantic Ocean*, *in* *The Troposphere of the Atlantic Ocean* by Albert Defant. U.S. Department of Commerce, Springfield, VA.
- England, M. 1995. The age of water and ventilation timescales in a global ocean model. *J. Phys. Oceanogr.*, *25*, 2756–2777.
- England, M. H., J. S. Godfrey, A. C. Hirst and M. Tomczak. 1993. The mechanism for Antarctic Intermediate Water renewal in a world ocean model. *J. Phys. Oceanogr.*, *23*, 1553–1560.
- England, M. H. and E. Maier-Reimer. 1999. Using chemical tracers in ocean models. *Rev. Geophys.* (submitted).
- Fedorov, K. N. 1978. *The Thermohaline Finestructure of the Ocean*, Pergamon, Oxford, 170 pp.
- 1986. *The Physical Nature and Structure of Oceanic Fronts*, Springer-Verlag, Berlin, 333 pp.
- Figueroa, H. A. 1996. World ocean density ratios. *J. Phys. Oceanogr.*, *26*, 267–275.
- Garrett, C. J. R. and W. H. Munk. 1975. Space-time scales of internal waves: a progress report. *J. Geophys. Res.*, *80*, 291–297.
- Gent, P. R., J. Willebrand, T. J. McDougall and J. C. McWilliams. 1995. Parameterizing eddy-induced tracer transports in ocean circulation models. *J. Phys. Oceanogr.*, *25*, 463–474.
- Godfrey, J. S. 1989. A Sverdrup model of the depth integrated flow for the world ocean allowing for island circulations. *Geophys. Astrophys. Fluid Dyn.*, *45*, 89–112.
- Haidvogel, D. B. and F. O. Bryan. 1992. Ocean general circulation modeling, *in* *Climate System Modeling*. K. E. Trenberth, ed., Cambridge University Press, 371–412.
- Ingham, M. C. 1966. *The salinity extrema of the world ocean*, PhD thesis, Oregon State University, Corvallis, OR.
- Jones, H. and J. Marshall. 1993. Convection with rotation: A study of open-ocean deep convection. *J. Phys. Oceanogr.*, *23*, 1009–1039.
- Kämpf, J. and J. O. Backhaus. 1998. Shallow, brine-driven free convection in polar oceans: nonhydrostatic numerical process studies. *J. Geophys. Res.*, *103*, 5577–5593.
- Karstensen, J. 1998. OMP (Optimum Multiparameter) analysis—User Group. [http://www.ifm.uni-hamburg.de/~wwwro/omp\\_std](http://www.ifm.uni-hamburg.de/~wwwro/omp_std)

- Karstensen, J. and M. Tomczak. 1998. Age determination of mixed water masses with CFC and oxygen data. *J. Geophys. Res.*, *103*, 18,599–18,610.
- Klein, B. and M. Tomczak. 1994. Identification of diapycnal mixing through optimum multiparameter analysis 2: Evidence for double-diffusive mixing in the front between North and South Atlantic Central Water. *J. Geophys. Res.*, *99*, 25,275–25,280.
- Lagerloef, G. S. E., C. T. Swift and D. M. Le Vine. 1995. Sea surface salinity: the next remote sensing challenge. *Oceanography*, *8*, 44–50.
- Mackas, D. L., K. L. Denman and A. F. Bennett. 1987. Least-square multiple tracer analysis of water mass composition. *J. Geophys. Res.*, *92*, 2907–2918.
- Mamayev, O. I. 1975. *Temperature-Salinity Analysis of World Ocean Waters*, Elsevier, Amsterdam.
- Marshall, J. C., A. J. G. Nurser and R. G. Williams. 1993. Inferring the subduction rate and period over the North Atlantic. *J. Phys. Oceanogr.*, *23*, 1315–1329.
- Masuzawa, J. 1969. Subtropical Mode Water. *Deep-Sea Res.*, *16*, 463–472.
- McCartney, M. 1977. Subantarctic Mode Water. *Deep-Sea Res.*, *24*, Suppl. (A voyage of Discovery), 103–119.
- McWilliams, J. C. 1985. Submesoscale, coherent vortices in the ocean. *Rev. Geophys.*, *23*, 165–182.
- Moiseyev, G. A. and A. G. Chertushkin. 1995. About some morphological characteristics of the oceanic fields' vertical structure. *Phys. Oceanogr.*, *6*, 377–387.
- Osborne, J., P. Rhines and J. Swift. 1991. OceanAtlas for Macintosh, a microcomputer application for examining oceanographic data, version 1.0. Scripps Institution of Oceanography ref. #91-5.
- Osborne, J. and J. Swift. 1996. Power OceanAtlas. International WOCE Newsletter 22, 43 (unpublished manuscript; see also <http://odf.ucsd.edu/OceanAtlas/>).
- Pickard, G. L., J. R. Donguy, C. Henin and F. Rougerie. 1977. A review of the physical oceanography of the Great Barrier Reef and western Coral Sea. Australian Institute of Marine Sciences Monograph Series Vol. 2, Australian Government Publishing Service.
- Poole, R. and M. Tomczak. 1999. Optimum multiparameter analysis of the water mass structure in the Atlantic Ocean thermocline. *Deep-Sea Res.* (in press).
- Putzka, A. 1995. Age distributions and transient tracer dating of water masses, *in* Application of Trace Substance Measurements to Oceanographic Problems, P. Schlosser, J. W. M. Smethie and W. S. Broecker, eds., Tucson, USA.
- Redfield, A. C., B. H. Ketchum and F. A. Richards. 1963. The influence of organisms on the composition of sea-water, *in* The Sea, M. N. Hill, ed., Vol. 2, 26–77.
- Reid, J. L. and R. J. Lynn. 1971. On the influence of the Norwegian-Greenland and Weddell seas upon the bottom waters of the Indian and Pacific oceans. *Deep-Sea Res.*, *18*, 1063–1088.
- Ribbe, J. and M. Tomczak. 1997. On convection and the formation of Subantarctic Mode Water in the Fine Resolution Antarctic Model (FRAM). *J. Mar. Syst.*, *13*, 137–154.
- Roether, W. 1994. Studying thermohaline circulation in the ocean by means of transient tracer data, *in* Ocean Processes in Climate Dynamics: Global and Mediterranean Examples, P. Malanotte-Rizzoli and A. R. Robinson, eds., 157–171.
- Roether, W., R. Schlitzer, A. Putzka, P. Beining, K. Bulsiewicz, G. Rohardt and F. Delahoyde. 1993. A chlorofluoromethane and hydrographic section across Drake Passage: Deep Water ventilation and meridional property transport. *J. Geophys. Res.*, *98*, 14,423–14,435.
- Rudels, B., D. Quadfasel, H. Friedrich and M.-N. Houssais. 1989. Greenland Sea convection in the winter of 1987–1988. *J. Geophys. Res.*, *94*, 3223–3227.
- Sankey, T. 1973. The formation of deep water in the northwestern Mediterranean. *Progr. Oceanogr.*, *6*, 159–188.
- Schmitt, R. 1981. Form of the temperature-salinity relationship in the central water: evidence for double-diffusive mixing. *J. Phys. Oceanogr.*, *11*, 1015–1026.

- 1990. On the density ratio balance in the central water. *J. Phys. Oceanogr.*, *20*, 900–906.
- Schott, F., M. Visbeck and J. Fischer. 1993. Observations of vertical currents and convection in the central Greenland Sea during the winter of 1988–1989. *J. Geophys. Res.*, *98*, 14,402–14,421.
- Send, U. and J. Marshall. 1995. Integral effects of deep convection. *J. Phys. Oceanogr.*, *25*, 855–872.
- Sprintall, J. and M. Tomczak. 1993. On the formation of Central Water and thermocline ventilation in the southern hemisphere. *Deep-Sea Res.*, *40*, 827–848.
- Stommel, H. 1979. Determination of water mass properties of water pumped down from the Ekman layer to the geostrophic flow below. *Proc. Nat. Acad. Sci. USA*, *76*, 3051–3055.
- Sverdrup, H. U., M. W. Johnson and R. H. Fleming. 1942. *The Oceans: Their Physics, Chemistry and General Biology*, Prentice-Hall, Englewood Cliffs, NJ, 1087 pp.
- Takahashi, T., W. S. Broecker and S. Langer. 1985. Redfield ratio based on chemical data from isopycnal surfaces. *J. Geophys. Res.*, *90*, 6907–6924.
- Tandon, A. and C. Garrett. 1996. On a recent parameterization of mesoscale eddies. *J. Phys. Oceanogr.*, *26*, 406–411.
- Thompson, R. O. R. Y. and W. Edwards. 1981. Mixing and water-mass formation in the Australian Subantarctic. *J. Phys. Oceanogr.*, *11*, 1399–1406.
- Tomczak, M. 1980. A review of Wüst's classification of the major deep-sea expeditions 1873–1960 and its extension to recent oceanographic research programs, *in* *Oceanography: The past*, M. Sears and D. Merriman, eds., Springer-Verlag, New York, 188–194.
- 1981a. A multi-parameter extension of T-S diagram techniques for the analysis of non-isopycnal mixing. *Prog. Oceanogr.*, *10*, 147–171.
- 1981b. An analysis of mixing in the frontal zone of South and North Atlantic Central Water off North-West Africa. *Prog. Oceanogr.*, *10*, 173–191.
- Tomczak, M. and Y. H. Gu. 1987. Water mass properties of the permanent thermocline in the western South Pacific Ocean during WESTROPAC '82. *Deep-Sea Res.*, *34*, 1713–1731.
- Tomczak, M. and D. G. B. Large. 1989. Optimum multiparameter analysis of mixing in the thermocline of the eastern Indian Ocean. *J. Geophys. Res.*, *94*, 16,141–16,149.
- Warren, B. A. 1983. Why is no deep water formed in the North Pacific? *J. Mar. Res.*, *41*, 327–347.
- 1992. Physical Oceanography in “The Oceans.” *Oceanography*, *5*, 157–159.
- Williams, R. G., M. A. Spall and J. C. Marshall. 1995. Does Stommel's mixed layer “demon” work? *J. Phys. Oceanogr.*, *25*, 3089–3102.
- Wüst, G. 1936. Schichtung und Zirkulation des Atlantischen Ozeans: Das Bodenwasser und die Gliederung der Atlantischen Tiefsee. *Wissenschaftliche Ergebnisse der Deutschen Atlantischen Expedition “Meteor” 1925–1927*, vol. VI part 1, 3–107.
- You, Y. and M. Tomczak. 1993. Thermocline circulation and ventilation in the Indian Ocean derived from water mass analysis. *Deep-Sea Res.*, *40*, 13–56.