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Salinity Patterns in the Ocean

Lynne D Talley
Scripps Institution of Oceanography, University of California, La Jolla, CA, USA

Ocean salinity varies geographically and with time. Fresh water input occurs at the sea surface due to precipitation and river inflow, reducing salinity. Salinity is increased by evaporation and also as a by-product of sea ice formation. Because freshening and salinification occur in different places, salinity at a particular location reflects the upstream source of the water there. In subtropical latitudes, high surface evaporation creates high salinity near the sea surface. In subpolar latitudes, high precipitation creates low salinity near the sea surface. As these waters flow into the ocean interior, they create layers of high and low salinity.

At mid-depth (i.e., around 1000 to 2000 m deep), outflows from the highly evaporative Mediterranean and Red Seas create a vertical salinity maximum in the North Atlantic and Indian Oceans, respectively. Also at mid-depth in the subtropical and tropical regions, the relatively fresh, but dense, surface water from higher latitudes flows in and creates a vertical salinity minimum, most prevalent in the Southern Hemisphere and North Pacific. The North Atlantic is the most saline ocean and the North Pacific the freshest.

Salinity affects sea water density and therefore can be a controlling factor for the depth of the ocean surface mixed layer. This depth also depends on wind speed and cooling at the sea surface. Feedbacks can occur as precipitation and evaporation are affected by the atmosphere, which in turn is affected by sea surface temperature. The atmosphere in turn is affected by sea surface temperature, which is affected by mixed layer depth.

The distribution of dissolved salts (*see* **Salinity**, Volume 1) in the oceans and adjacent seas varies in space and time. Salinity is changed near the sea surface by precipitation and evaporation of fresh water and by salty water produced sea ice forms and excludes salt. Geographical variations in inputs create regional differences in salinity at the sea surface. As seawater circulates down into the ocean away from the sea surface, it carries these differences along, creating large-scale salinity patterns throughout the ocean. Changes in time in the inputs at the sea surface also affect the salinity distribution. By mapping the salinity distribution and tracking changes in salinity over time, much insight is gained into both the basic ocean circulation and the processes that change the ocean circulation and temperature distribution.

Salinity is a seawater property related to the amount of matter, mainly consisting of salts, dissolved in the water (*see* **Salinity**, Volume 1). The original definition of salinity was in terms of grams of dissolved salts per kilogram of seawater. Salinity is now defined in terms of seawater conductivity, and is currently calibrated using a standard solution prepared at a laboratory in the UK. The quantity of dissolved salts, hence the salinity, affects seawater density, which also depends on temperature and pressure. Quantifying these dissolved constituents as the single parameter, salinity, is possible because the source of solids is weathering, which occurs on geological time scales. Ocean circulation and mixing, on the other hand, distribute the solids throughout the World Ocean on a time scale of no more than thousands of years, and so the various salt constituents are essentially well mixed. Therefore, salinity differences are created only by dilution or concentration as fresh water is added or removed, or as salty water is rejected from sea ice as it freezes. Other dissolved non-salt constituents such as carbon, calcium and silica, which are affected by biological cycles, have enough spatial variability to affect seawater density (e.g., Millero *et al.*, 1978); this small variation is not usually included in salinity and density studies.

In many ocean regions, particularly in the tropics, the subtropics and the deep ocean, temperature variations are more important than salinity in changing density. However, at higher latitudes, the presence of freshened surface waters resulting from high precipitation and/or ice melt has a strong influence on whether the water column can be cooled enough to mix vertically with other layers. Because winter surface cooling at high-latitudes determines water properties for the global ocean at depths below about 1500 m, salinity can control the occurrence of convection and its penetration depth. With a surface layer that is relatively fresh compared with the underlying water that came from another ocean region, surface waters can cool to freezing without overturning. If the high-latitude surface waters are not much freshened, as is the usual case in the northern North Atlantic, then overturn to great depth at the end of winter can occur with even modest surface cooling, leaving the surface waters well above the freezing point. Because ocean surface temperatures are part of the forcing for the atmosphere, changes in high-latitude salinity distribution can be involved in climate change.

In addition to its effect on seawater density, salinity is useful in itself as a tracer of ocean circulation and mixing. Salinity is established in the ocean's surface layer by local processes, and therefore salinity variations within the ocean depend on the surface origin of a parcel of water. Mixing between waters of different salinities occurs within the ocean, and so the original salinity changes gradually. However, mixing rates are low enough that salinity (and other chemical tracers) can provide a great deal of information on the pathways of flow.

SEA SURFACE FORCING OF SALINITY: PRECIPITATION, EVAPORATION, RUNOFF AND ICE PROCESSES

The total amount of salt in the ocean is constant on time scales of millions of years, and hence is assumed constant for studies at almost all time scales. However, the total amount of water in the oceans varies daily due to exchange at the sea surface. Rainfall, runoff from land, and ice melt all reduce salinity by increasing the amount of fresh water. Evaporation increases salinity by decreasing the amount of fresh water.

Ice formation also locally increases salinity through a process called brine rejection. As ice forms from seawater, the dissolved salts are rejected from most of the ice and collect in pockets in the ice. This briny water drips through the base of the forming ice. More brine is released during slow ice formation than if the ice forms quickly, so the strength of brine rejection depends on the local air temperature. The brine mixes with the seawater beneath the ice, raising the seawater's salinity. The net salinity increase depends on the thickness of the mixing layer below the ice. In shallow locations, such as on shallow continental shelves, salinity might increase more than over a deep bottom where mixing occurs to greater depth. Because the water below sea ice is usually very cold, the addition of salty brine may increase salinity without increasing temperature. This results in a density increase in the seawater. This denser water then sinks until it reaches a depth where surrounding waters have the same density. Because these deeper waters may have come, at least in part, from somewhere else (e.g., in the Arctic from the North Atlantic Ocean), these waters will usually not be as cold as the new ice-rejected brine. Although warmer, their density will be equal to the ice-rejected brine because they are even saltier. That is because the ice-rejected brine was formed from near-freezing surface waters that initially had much lower salinity (e.g., because they had a higher proportion of meltwater and river runoff), and therefore lower density, than the deeper waters. Although it seems paradoxical, the ice formation and brine rejection processes are thus able to create the coldest and freshest waters on a surface of constant density within the ocean.

Because sea ice has lower salinity than the original seawater from which it formed, melting of sea ice creates a fresh surface layer of low-density waters that float over the more saline waters below.

Ocean salinity patterns at the largest scales are affected primarily by the annual average of evaporation and precipitation, with additional major effects from ice melt and freezing at high-latitudes. Evaporation and precipitation, usually reported in cm year^{-1}, are primarily calculated from observations collected on a routine basis by most ships. The difference between evaporation (E) and precipitation (P) has been calculated from the complete historical data set compiled from these ship observations (Figure 1). The range of E−P is about −10 to 10 cm year^{-1}. There is more evaporation than precipitation in the subtropical regions between about 15° and 40 °S and N latitude, under the atmosphere's large-scale high pressure centers, with descending, dry air. At higher latitudes, there is more precipitation than evaporation, under the atmospheric low-pressure centers. In the tropics, especially at about 10 °N over the Pacific and Atlantic, surface air rises in the intertropical convergence zone (ITCZ), forming enormous cumulus clouds that produce large amounts of rain.

Runoff from land affects the coastal salinity distribution. On a global scale, runoff is most important at the major river outlets, indicated in Figure 1. The net fresh water input is not shown in Figure 1, but the effect of these major rivers is apparent in the surface salinity distribution considered next.

SURFACE SALINITY DISTRIBUTION

Salinity at the sea surface (Figure 2) resembles the evaporation−precipitation pattern, with some differences due to runoff, ocean currents and ice melt. The total salinity range of the open ocean is about 30−40 parts per thousand or ‰ (*see* **Salinity**, Volume 1). The international convention is that salinity has no units; its value is approximately grams salt per kilogram of sea water. Many authors continue to use the practical salinity unit (psu) rather than leaving the quantity unitless. While salinity in coastal estuaries can be much lower, it does not affect large oceanic-scale patterns. Lowest salinities occur in the Arctic and Antarctic where there is both net precipitation and seasonal ice melt. Highest salinities occur in the Red Sea and Persian Gulf, both located in the northwestern Indian Ocean, where net evaporation is high. High salinity is also found in the Mediterranean Sea. In the open ocean, high salinity occurs in the subtropical areas of net evaporation seen in Figure 1. A band of low salinity underlies the ITCZ at 10 °N.

The effect of continental runoff is apparent in lowered surface salinity near the mouths of major rivers such as the Amazon and the Congo and the numerous large rivers that empty into the Bay of Bengal, east of India, including the Ganges and Brahmaputra. Runoff from many rivers around the Gulf of Alaska in the northeastern Pacific and around the Arctic Ocean is important in the lowered salinities of high-latitude ocean regions.

VERTICAL STRUCTURE OF THE SALINITY DISTRIBUTION

Waters at the sea surface flow down into the interior. Salinities from the sea surface (Figure 2) carried with this flow into the ocean interior are helpful in identifying sources and pathways of the circulation. The two primary

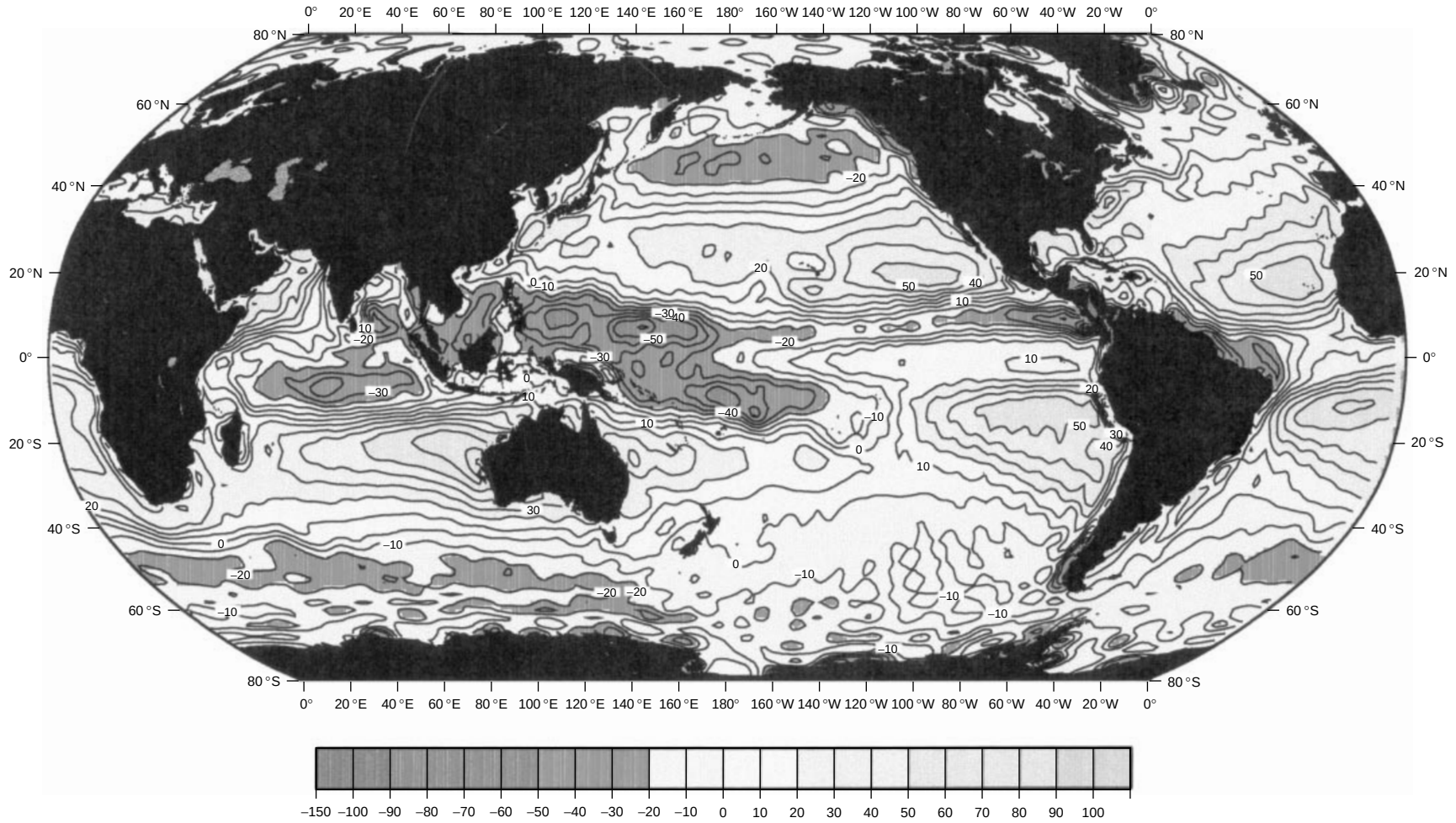


Figure 1 Evaporation minus precipitation (cm year^{-1}), annual average. The major rivers of the world are also indicated. (Reproduced using data from the National Centers for Environmental Prediction Reanalysis Project, Kalnay *et al.*, 1996)

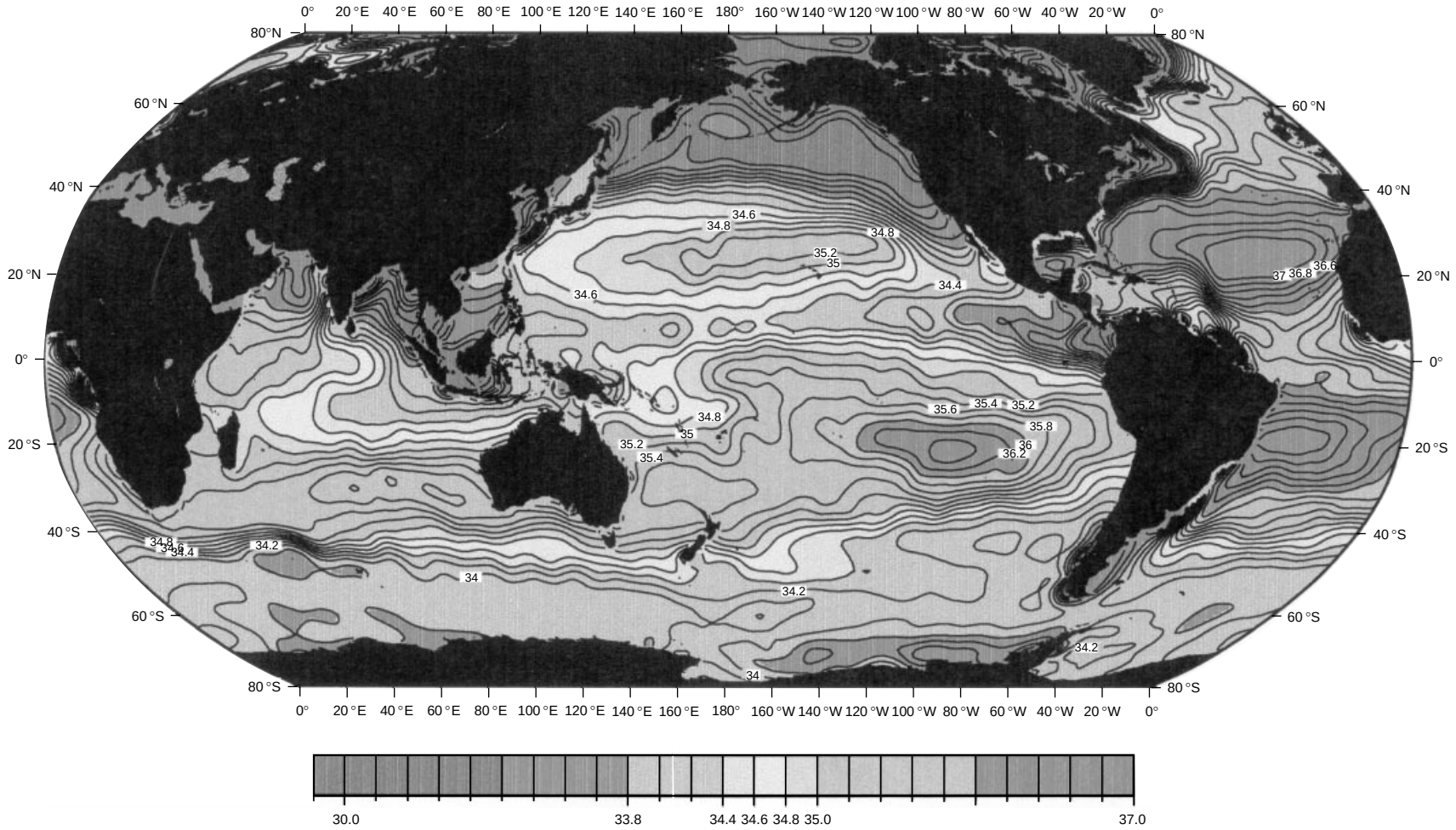


Figure 2 Surface salinity (g salt per kg of seawater), using annually averaged ship observations. (Reproduced using data from Levitus *et al.*, 1994)

mechanisms for this downward flow are: (1) flow following constant density surfaces that slope gently into the ocean interior, and (2) vertical convection and brine rejection at higher latitudes.

Flow down along constant density surfaces is primarily through a process called *subduction*, occurring in subtropical regions between about 15° and 40 °S and N latitude. In this process, waters in the surface mixed layer circulate equatorward, encountering warmer, less dense waters to the south, which they are carried beneath. Surface waters are most saline in the middle of these subtropical regions. Therefore, all equatorward subtropical regions are characterized by a subsurface vertical maximum in salinity caused by subduction of more saline waters from the center of the gyre, as described later (*see* **Ocean Circulation**, Volume 1).

Convection occurs in many places, but reaches most deeply into the ocean at high latitudes. Convection is not so much a downdraft of water from the surface as it is a deep mixing, which brings surface water to great depth, from whence the water spreads laterally along constant density surfaces. Major deep convection areas can be identified from subsurface properties, such as the salinity distributions discussed below. In the North Atlantic and Arctic, these sites are in the Labrador Sea between Labrador and Greenland, in the Greenland Sea east of Greenland and north of Iceland, and in the Mediterranean Sea. In the North Pacific, convection in the Okhotsk and Japan Seas is significant in modifying water properties at mid-depth. In the Southern Hemisphere, convection occurs all around Antarctica at about 40° to 45 °S, which is just north of the major current called the Antarctic circumpolar current. Brine rejection is important for subsurface water properties primarily in the Antarctic (Weddell and Ross Seas and other locations around the continent), the Okhotsk Sea, and the interior of the Arctic Ocean.

The following definitions are useful for discussing the salinity distribution in the vertical: (1) *halocline* – where the salinity value changes rapidly in the vertical in comparison with changes above and below the halocline; (2) *halostad* – where salinity changes very slowly in the vertical; (3) *intermediate water* – usually referring to a minimum or maximum in salinity in the vertical which is found over a large horizontal area at mid-depth in the ocean. Equivalent terms for vertical variations in temperature and density are *thermocline*, *thermostat*, *pycnocline* and *pycnostad*.

SUBTROPICAL SALINITY STRUCTURE

In the subtropical regions of every ocean, salinity (Figures 3 and 4a) is high close to the sea surface due to strong evaporation. At the location of the highest surface salinity (Figure 2), the highest salinity in vertical profiles is at the sea surface. This high salinity water is carried downward and equatorward through subduction. Therefore, equatorward of the center of the highest surface salinity, there is a vertical maximum in salinity. This salinity maximum is at about 50 m depth, which is very shallow compared with the ocean depth. These subsurface salinity maxima are sometimes called *Subtropical Underwater* (Worthington, 1976).

At all locations in the subtropical gyres, and below the near-surface salinity maximum, salinity decreases smoothly with depth to a salinity minimum. Temperature also decreases smoothly and rapidly (Figure 4b). This layer of changing properties is called the *thermocline* (region of high temperature change in the vertical), and is also often called *central water*. Central water is the subducted water of the subtropical gyre; the colder waters come from farther north within the subtropical circulation, with little input from the subpolar region. It is clear from Figure 4(b) that the North Atlantic central waters are the most saline and the North Pacific the freshest.

Below the central waters, at depths between about 500 and 1500 m, each subtropical gyre (except in the North Atlantic) has a vertical minimum in salinity. The salinity minimum in the North Pacific is called *North Pacific intermediate water* (Reid, 1965). It arises from cold, low salinity surface waters in the subpolar region of the North Pacific, specifically in the Okhotsk Sea and near the Kuril Islands (Talley, 1991, 1993). These waters flow southward along the western boundary in the Oyashio current and enter the subtropical circulation east of Japan. The existence of the low salinity layer throughout the subtropical circulation reflects the presence of this subpolar water underlying the waters from subtropical sources.

The salinity minimum in the Southern Hemisphere is called *Antarctic intermediate water* (Reid, 1965). It comes from low salinity surface waters found just west of Chile and just east of Argentina (McCartney, 1977). Antarctic intermediate water in the Pacific Ocean originates primarily from the Chilean source. The Atlantic and Indian Ocean Antarctic intermediate waters originate from the Argentinean source. The Argentinean waters arise from Chilean surface waters, which flow eastward through the northern side of Drake Passage between South America and Antarctica. The Chilean waters are modified somewhat during their passage into the Atlantic. Therefore, the properties of the Atlantic and Indian Antarctic intermediate waters differ slightly from those of the Pacific Antarctic intermediate water.

The salinity structure in the subtropical North Atlantic is somewhat different. While there is a salinity minimum at the base of the central water, this results primarily from the northernmost subducted waters of the subtropical region, and is not the major fresh intermediate water of

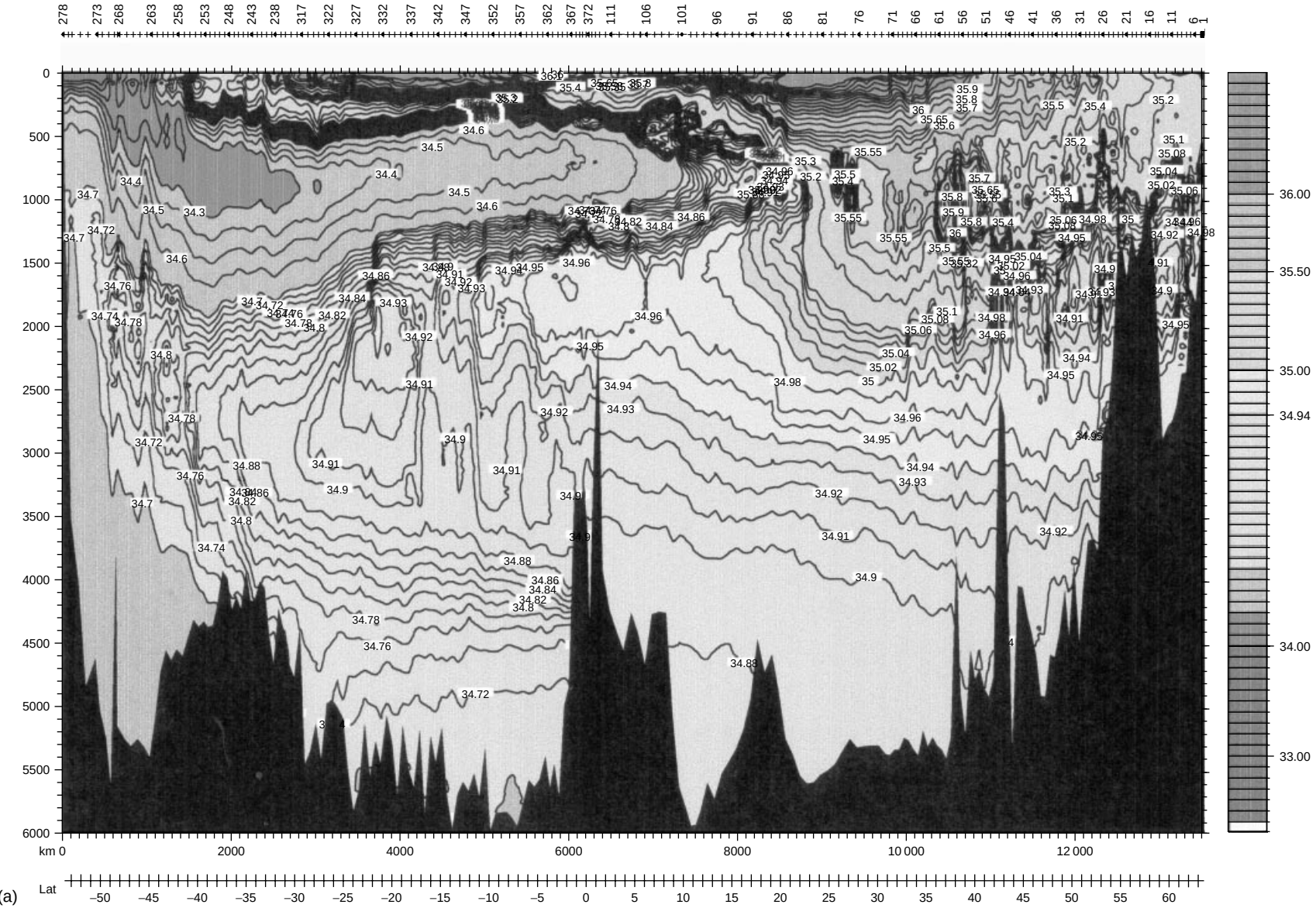


Figure 3 (a) Vertical cross-section of the Atlantic Ocean, along about 20 °W. (b) Vertical cross-section of the Pacific Ocean, along about 150 °W. (c) Vertical cross-section of the Indian Ocean, along about 90 °E. (Reproduced using data from the World Ocean Circulation Experiment)

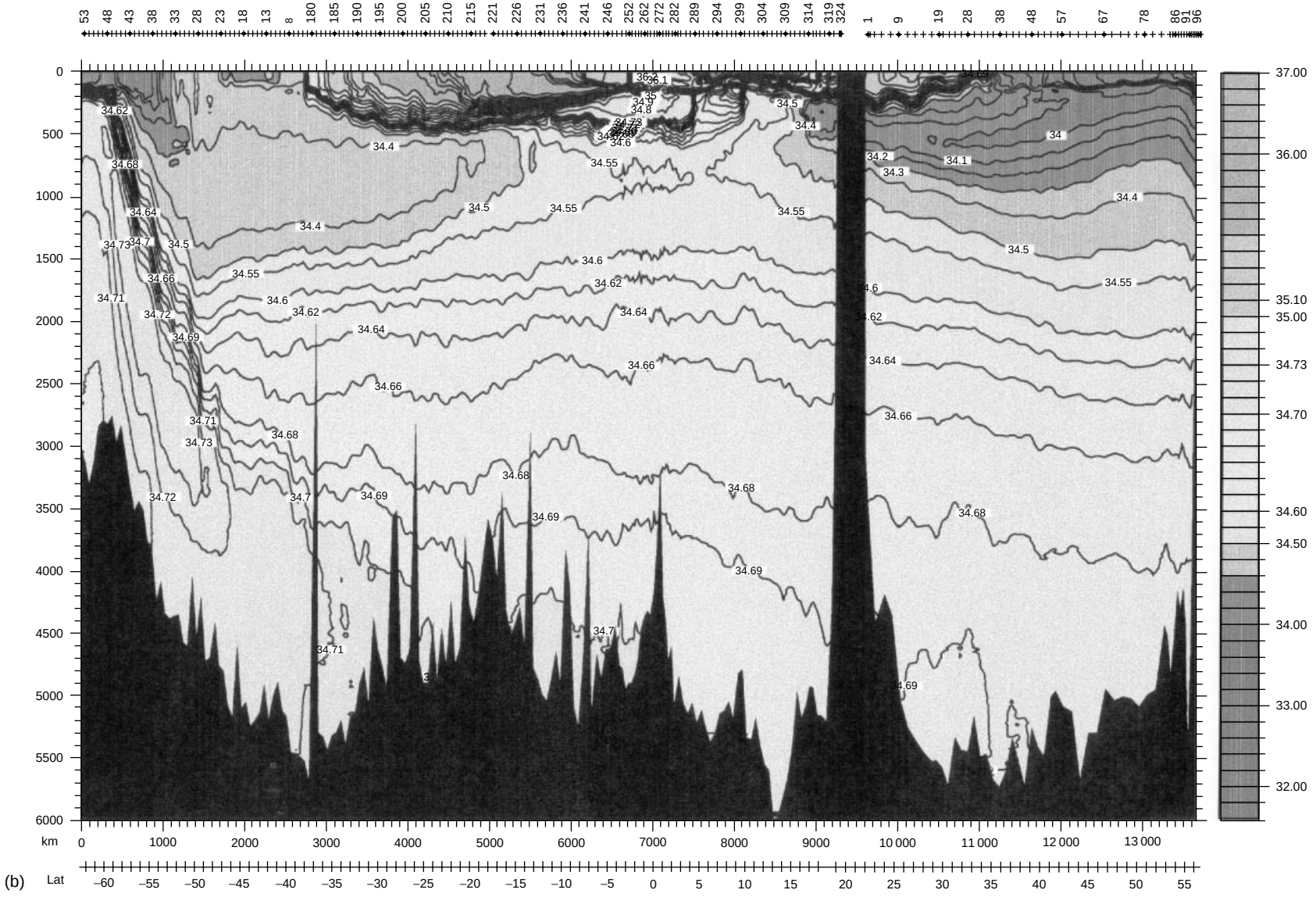
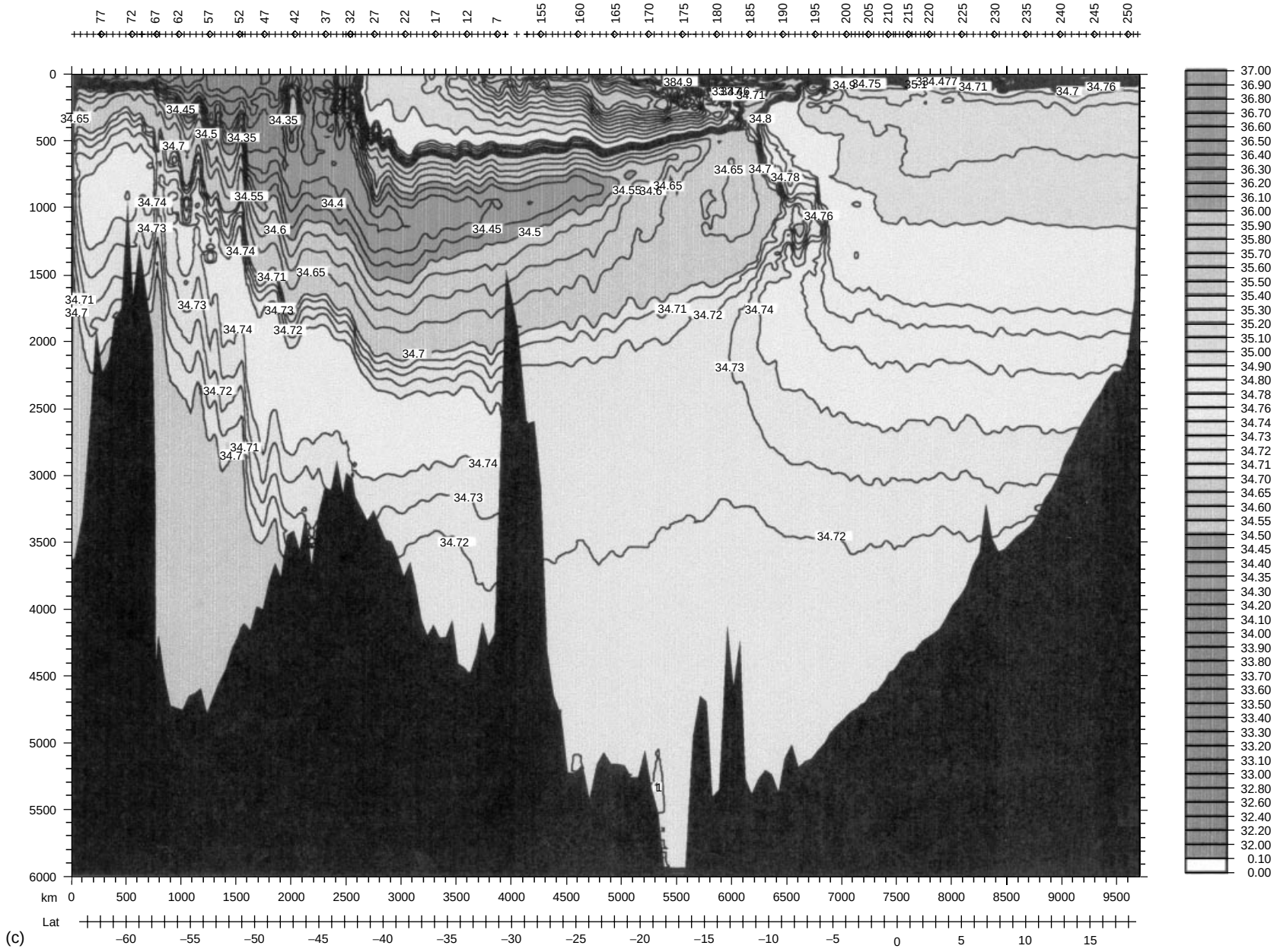


Figure 3 (*Continued*)



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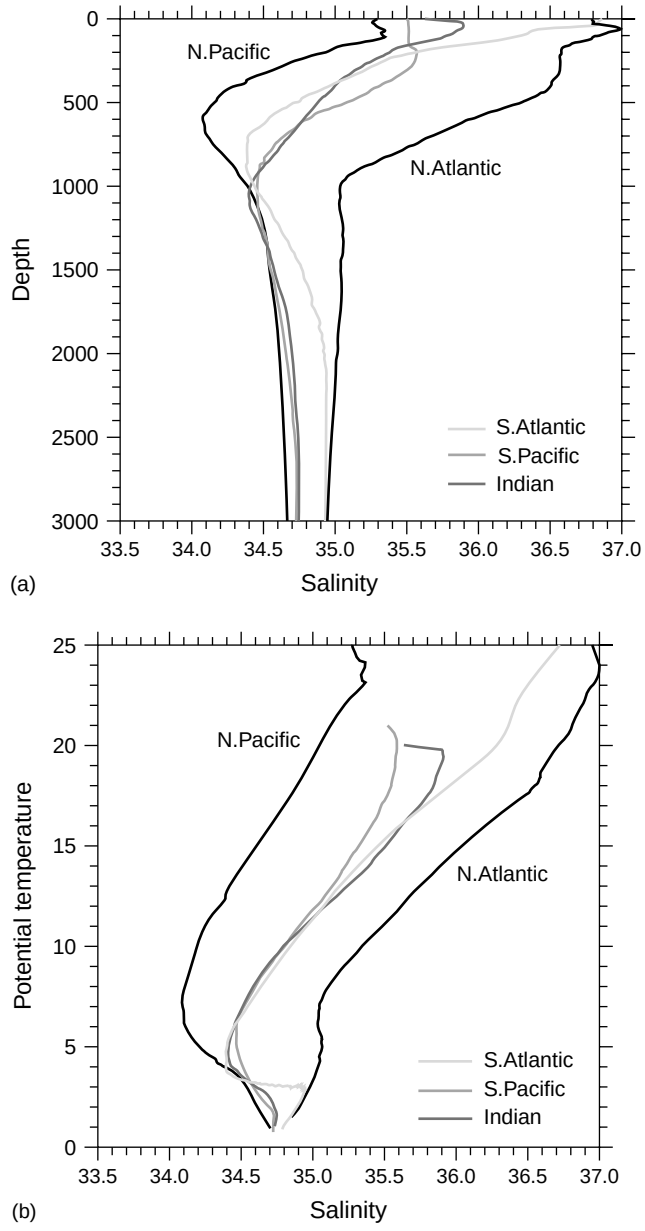

Figure 4 (a) Salinity distribution with depth from each subtropical gyre. All profiles are chosen from the centers of the subtropical circulations, at about 25°N (Northern Hemisphere) and 30°S (Southern Hemisphere), and from the center of the basins rather than near the boundaries. (b) Potential temperature as a function of salinity for each subtropical ocean basin

the North Atlantic, which dominates the subpolar region (described later). The main fresh intermediate water is displaced in the subtropical gyre by the high salinity outflow at mid-depth from the Mediterranean Sea. This outflow creates an intermediate depth salinity maximum in the vertical.

In the Indian Ocean, a dense, high salinity source in the Red Sea creates a high-salinity intermediate water in the north, similar to the Mediterranean water of the Atlantic. Because there are no northern high-latitude sources of water in the Indian Ocean, there is no fresh intermediate layer arising from the north. The fresh water at the surface in the Bay of Bengal arising from major river runoff is much too warm to sink to mid-depth. Thus, the mid-depth northern Indian Ocean is dominated by the highly saline outflow from the Red Sea.

Beneath the low salinity intermediate waters of the North Pacific and Southern Hemisphere, salinity increases towards the bottom. The higher salinity of the deep waters of the Southern Hemisphere and North Pacific results from input of saline North Atlantic waters. Beneath the saline intermediate water of the North Atlantic and tropical Indian Oceans, salinity decreases to the bottom. In the South Atlantic Ocean, low-salinity intermediate water (Antarctic intermediate water) lies above higher salinity deep waters from the North Atlantic. Beneath the high-salinity North Atlantic deep water, salinity decreases to the bottom. The lower salinity of the bottom waters in the Atlantic and Indian Oceans results from deepwater sources in the Antarctic that are of lower salinity.

SUBPOLAR SALINITY STRUCTURE

In high latitude regions dominated by precipitation and ice melt, salinity is lowest at the sea surface and increases downward through a halocline (Figure 5, North and South

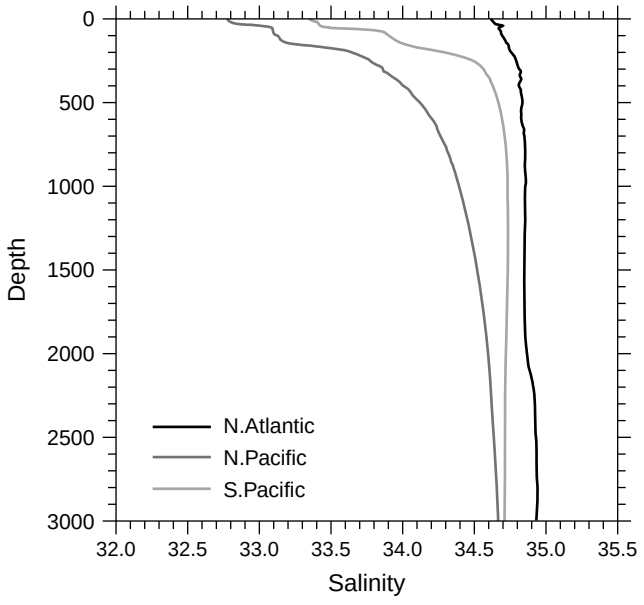

Figure 5 Salinity distribution with depth from the subpolar regions. The Northern Hemisphere profiles are from about 50°N. The Southern Hemisphere profile is from the Pacific Ocean at about 65°S

Pacific profiles). In the North Pacific and Southern Hemisphere, and also in most of the Arctic Ocean, the halocline is very sharp, with a large change in salinity over several tens of meters. Because of the salinity structure, the surface layer can be colder than the water below the halocline. This pre-conditions the region for sea ice formation. The cooled surface waters can circulate away from their winter outcrop locations and appear as a subsurface temperature minimum between warmer water below from inflow from other regions and warmer water above from local seasonal warming.

The North Atlantic's subpolar gyre is affected by inflow of a thick layer of saline, warm surface water from the subtropical gyre. Therefore, the halocline typical of the subpolar Pacific and subpolar Southern Hemisphere is absent in most of the subpolar North Atlantic (Figure 5, North Atlantic profile). The saline flow into the central and eastern subpolar gyre is affected by precipitation, and the surface waters become gradually fresher as they circulate counterclockwise. When the surface waters enter the Labrador Sea, they are much fresher than when they first enter the subpolar region. Winter convection to about 1500 m depth in the Labrador Sea then mixes the relatively fresh water to intermediate depth. This spreads away from the Labrador Sea and forms a salinity minimum throughout the subpolar gyre, beneath the inflowing warm, saline waters. This differs from subpolar regions in the other oceans where the surface waters are strongly dominated by local fresh water sources. Therefore, most of the North Atlantic's subpolar gyre has a vertical salinity structure similar to a subtropical structure, with saline waters at the surface and an intermediate depth salinity minimum.

TROPICAL SALINITY STRUCTURE

The upper ocean salinity distribution depends on local precipitation patterns. Thus, a band of low salinity is found under the high precipitation of the ITCZ at 5–10 °N in the Pacific. This band is separated from the saline waters below it by a halocline. Excess precipitation is also found in the western Pacific at the equator, creating a halocline there as well. The eastern equatorial Pacific on the other hand is dominated by evaporation and so there is no halocline there. The Bay of Bengal in the Indian Ocean is the site of major runoff from Asia; this relatively fresh, warm water creates a shallow halocline throughout the northeast Indian Ocean. The intermediate and deepwater salinity distributions in each ocean within 15° of the equator are similar to those of the adjacent subtropical regions.

GLOBAL SCALE SALINITY PATTERNS

As indicated above, there are significant differences in salinity between oceans. Considering the total ocean basin, the North Atlantic is the most saline and the North Pacific the freshest of the major oceans. Because of significant river inflows, the surface waters of the Arctic are the freshest. The three Southern Hemisphere oceans are connected around Antarctica and hence their salinities are similar, midway between the North Atlantic and North Pacific. The main reason for the differences between the North Pacific and North Atlantic salinities is the difference in net evaporation per unit area. There is relatively more evaporation in the North Atlantic's subtropical regions than in the North Pacific's subtropics. The evaporative regions occur where there are easterly trade winds. The North Atlantic's trade winds originate over the deserts of Africa and the Middle-East. The North Pacific's trade winds originate over the much smaller continental areas of Central America and Mexico. Therefore, the North Atlantic's trade winds are drier than the North Pacific's and cause more evaporation per unit area. High salinity in the North Atlantic is present at intermediate depth because of the Mediterranean Sea. The North Atlantic exchanges water with the Mediterranean Sea through the Strait of Gibraltar. Evaporation and cooling in the Mediterranean cause the outflowing water to be much denser than the inflow. This dense outflow sinks to intermediate depth and creates the mid-depth high salinity that marks North Atlantic waters. This input of high salinity at mid-depth in the North Atlantic is one reason for the overall higher salinity of the North Atlantic compared with the North Pacific. Similarly, in the Indian Ocean, high evaporation in the Arabian region creates very saline and dense water in the Red Sea, which enters the Indian Ocean and raises the salinity at mid-depths compared with the inflowing fresher deep waters from the Antarctic.

ROLE OF SALINITY IN CLIMATE CHANGE

Salinity is a major factor in climate change because it can affect the depth of mixing at the sea surface. If for instance there is a shallow, fresh layer atop a denser, more saline layer, even large surface cooling can act only on the surface layer, creating a large temperature drop without deep convection. In the absence of a halocline, heat can be removed from a much thicker layer, which then results in a smaller temperature change. The processes that set the salinity of the waters beneath the surface layer, and hence the strength of the halocline, are complex. This salinity control of surface-layer processes in the present climate has been studied, e.g., in the tropical Pacific, as part of the El Niño and La Niña cycle. In the normal state in the northern Atlantic, up to and including the region between Greenland and Norway, surface salinities are not much lower than those of the underlying waters, permitting overturn to more than several hundred meters depth. This overturn is part of the global thermohaline circulation in which deep waters

are produced from surface waters at high-latitudes, mainly through heat loss to the atmosphere. (The presence of a permanent and strong halocline in the northern North Pacific and in the Arctic Ocean prevents deep overturning of waters there.) During observed decadal climate changes in the North Atlantic, a halocline sometimes forms as a result of excessive ice melt and runoff and is associated with shutdown of deep wintertime convection. This then affects the sea surface temperature as described previously, and can then affect the atmosphere. From ocean computer modeling experiments it is hypothesized that similar controls by salinity on deep overturn have occurred on much longer climate timescales.

See also: **Broecker, Wallace S**, Volume 1; **Ocean Circulation**, Volume 1.

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