Title
On the World-wide Circulation of the Deeper Waters of the World Ocean

Author
Reid, Joseph L

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On the world-wide circulation of the deeper waters of the World Ocean

Joseph L. Reid

Scripps Institution of Oceanography

Although the large-scale circulation of the surface waters of the world ocean has been revealed by the observations of mariners, the deeper patterns have not been so clearly defined. Therefore it seems worthwhile to describe the flow of some of the layers that can be recognized by their characteristics as they flow and spread through the ocean along various deeper paths. This work might be considered an introduction to a first course on physical oceanography.

As a framework we assume that the major driving forces of the ocean circulation are the winds and the exchange of heat and water through the sea surface. We know that the ocean is stratified in temperature, salinity, and density, and we assume that the ocean is nearly in geostrophic balance.

The Upper Waters

To show the shape and characteristics of the ocean we look first at the Pacific.

The major winds (fig. 1) are the Northeast and Southeast Trades, which blow westward in low latitudes, but with weaker winds in between, and the westerlies which blow eastward in high latitudes. These cause great gyres in mid-ocean (fig. 2). They are anticyclonic in mid-latitudes and cyclonic in high latitudes. Heat and water exchange through the sea surface establishes the surface density. This leads to overturn and lateral extension of the surface characteristics throughout the deep ocean.

The surface temperature is low in high latitudes, and high near the equator (fig. 3). The temperature decreases downward, and the isotherms are not horizontal, but slope (fig. 4). Salinity (fig. 5) is high near the tropic circles, where evaporation is greater than rainfall, and low where rainfall dominates in high latitudes and near the equator. It varies with depth fig. 6), and shows layers of low salinity from the surface extending equatorward beneath the surface.

Density and Geostrophy

Density (fig. 7) is labeled here in units of density anomaly ρ-1 where density (ρ) is expressed in kg/m^3. It decreases with rising temperature, but increases with increasing salinity and pressure. At the sea surface it is lowest in low latitudes and highest in the far north and south.

Water is compressible and if a parcel is moved up or down adiabatically its temperature falls or rises slightly. To compare two parcels at different depths it is convenient to calculate the temperature they would have if moved adiabatically to the sea surface, and call it the potential temperature. We can calculate the depth of isopycnals, along which water can move and mix laterally without altering the density structure.

Density increases with pressure (fig. 8). The isopycnals slope and it is convenient to compare the density of two parcels by referring their density to a common pressure, called potential density. The reference pressures we use here are zero, 1000, 2000, 3000, 4000, and 5000 decibars, and we call the density σ_0, σ_1, σ_2, σ_3, σ_4, and σ_5 according to the pressure of the parcel.

The ocean is stratified and the density varies horizontally as well as vertically, with the surface density lowest near the equator and increasing to the highest in high latitudes. The density also varies horizontally beneath the surface. There are horizontal density gradients at all depths between low-density and high-density areas. We see from the density field that the isopycnals that lie shallow or outcrop in high latitudes lie deeper in mid-latitudes.

The height of the sea surface above a deep pressure surface can be calculated from the measured salinity and temperature fields. This is the steric height which is $\int_{p2}^{p1} \propto d\rho$, where $\propto$ is he specific volume and ρ is pressure.

We find that the sea surface stands highest in mid-latitudes (fig. 9), along the axes of the great anticyclonic gyres, and lowest in high latitudes, along the axes of the cyclonic gyres.

We observe that the sea surface is not level but slopes upward and downward, with ridges and troughs. From the ridges to the troughs there must be a down-hill acceleration of $g(\partial z/\partial x)$, where g is gravity and z and x are the vertical and horizontal coordinates.

If there were no process opposing these pressure gradients the density structure we see would collapse, with some parts sinking and others rising to make the density field horizontally uniform.

But the earth rotates once in 24 hours and any object in motion upon it is subject to a Coriolis acceleration with respect to the earth, of $2\omega \sin \varphi\upsilon$ (abbreviated to $f\upsilon$) where ω is the earth's rotation, φ is the latitude, and υ is the velocity. This acceleration is to the right in the northern hemisphere and to the left in the southern hemisphere.

This leads to a transfer of some of the flow to the right—the Ekman transport—in the northern hemisphere and to the left in the southern hemispheres and a convergence of surface waters in the zone between the Trade Winds and the Westerlies, and into the anticyclonic gyres whose surface stands higher than the surrounding waters, and slopes downward from the center. Likewise the centers of the cyclonic gyres of high latitudes stand lower than the surrounding waters because the Ekman transport moves some of the flow away from the axes.

Where the slope of the surface causing an acceleration $g(\partial z/\partial x)$ is balanced by an equal and opposite acceleration $f\upsilon$, then a steady flow υ can be maintained. It is from these slopes that we can calculate the geostrophic flow. We cannot measure the slope directly, but if we know the temperature and salinity we can calculate the height of one surface p_1 with respect to another p_2. We get this from the steric height (ψ), which is $\int_{p1}^{p2} \propto dp$ where $\propto$ is the specific volume and p the pressure.

The velocity υ_g, then, is $\partial\psi/\partial x = \upsilon_1 - \upsilon_2 = \upsilon_g$, the flow along p_1 relative to p_2.

However, if we know something about the deeper patterns of flow we do not have to use a pressure surface to calculate the slopes, but can use the estimated or measured deep flow as a reference and get a much better picture of the steric height of the ocean, and can calculate the flow at all depths.

The slope of the sea surface defines the horizontal prssure gradient and the geostrophic flow at the surface, but beneath the surface it is the slope of the isopycnals that defines the pressure gradients, which are balanced by the earth's rotation. As these slopes vary both horizontally and vertically, the geostrophic flow also varies horizontally and vertically.

On a line of stations from Antarctica to Alaska in mid-ocean (fig. 9) we see the rise and fall of the sea surface that balances the cyclonic and anticyclonic flow patterns.

The ocean stands highest within the great anticyclonic gyres and lowest within the cyclonic gyres of high latitudes

The Coriolis acceleration for a given speed is $2\omega \sin \varphi \upsilon$ and is higher in high latitudes and the opposing slope for a given speed is steeper. The Pacific Ocean extends much farther poleward in the south than in the north, and the slope across the eastward-flowing Antarctic Circumpolar Current extends the decrease in steric height. Therefore the surface of the ocean stands lowest in the far South Pacific.

In fig. 10 the steric height of the surface of the Pacific Ocean has been adjusted to include the effect of the deep flow. It shows the effect of the Trade Winds and the Westerlies. At the continents these zonal flows turn back and form large gyres.

Where the wind stress and flow are in the same direction, as along the equator and in the North and South Equatorial currents near about 15°N and 15°S, the steric height rises along the flow, and the westward flow at the surface is slightly up hill. Where they are opposed, as in the equatorial countercurrent, about 5°N, the steric height will dip and the surface flow is slightly downhill.

The sea surface stands highest along the axes of the two largest gyres, with their axes near the tropic circles. They both turn anticyclonically, which is clockwise in the northern hemisphere and counterclockwise in the southern hemisphere.

Poleward of the anticyclonic gyres there are cyclonic gyres, centered near 50° to 70° latitude. There is a strong eastward countercurrent near between about 5°N and 10°N and a weaker equatorial countercurrent near 5°S in the west.

The flows are stronger along the western boundaries of the anticyclonic gyres, just east of Japan and Australia, and the slopes are steeper (fig. 10).

<u>Spreading beneath the surface</u>

The density at the sea surface (fig. 7) is highest in high latitudes, where the denser waters outcrop (fig. 8), in accord with the geostrophic balance.

The temperature is lowest in high latitudes and highest between the tropic circles (fig. 3). Cooler waters extend equatorward along the eastern boundaries, corresponding to the great anticyclonic gyres.

The surface salinity (fig. 5) is highest near the tropic circles, between the zones of highest rainfall along the equator and in the highest latitudes. This is important because we can use the salinity to trace the spreading of the water from the surface layer to subsurface depths.

In the stratified ocean, which does not overturn to the bottom everywhere, how are the surface characteristics transmitted downward? Do the subsurface patterns show only some vertical mixing and flow, or is something else going on?

We see from the potential density along the north-south section (fig. 8) that some of the isopycnals outcrop.

Some of the shallower isopycnals outcrop at both ends. Some of the denser isoycnals outcrop only in the south, and some do not outcrop at all in the Pacific.

We see that in addition to some vertical mixing in the surface layer there is also a spreading of these characteristics laterally from their source at the surface (fig. 6). That is, not just horizontally, but also laterally, along isopycnals.

Why along isopycnals? The simplest explanation is that such spreading requires a minimum of work. Along an isopycnal one parcel can replace another parcel without disturbing the density field. A parcel of cold and low-salinity water may have the same density as one that is warm and saline, and these two can mix freely.

Vertical mixing requires a stirring, a lifting of denser water into less dense water, or vice versa, and the density stratification of the ocean tends to limit this process in most areas of the ocean. But lateral mixing does not require such work against gravity. The layers simply float through the ocean at the depths where their density is found, and their depth varies in space in accord with the density structure.

We see the upper layer in mid-latitudes are made high in salinity by excess evaporation. Because of the geostrophically-balanced density structure, the isopycnals that outcrop there extend equatorward beneath the surface, and the high salinity from the surface waters might be expected to extend equatorward along that density range, just as a layer of orange dye.

Likewise, in high latitudes the surface waters are made low in salinity by excess precipitation, and the isopycnals that outcrop there also extend toward the equator beneath the surface. Low salinities may extend subsurface toward the equator from both northern and southern high latitudes just as a layer of blue dye would extend.

If we look at a vertical section of salinity along the middle of the Pacific (fig.6) and at density (fig. 8), we see that patterns of salinity seem to follow the slopes of the isopycnals.

For temperature (fig. 4) we find a monotonic decrease with increasing depth over most of the section, but we see warm water extending deeper in mid-latitudes where the isopycnals lie deeper than in high and low latitudes.

Salinity (fig. 6) gives the clearest picture of these isopycnal extensions, with tongues of high and low salinity extending from their sources along the isopycnals that outcrop in the areas of highest and lowest surface salinity.

We can pick out an isopycnal from the density section (fig. 8) that lies within the shallow salinity minimum and look at its depth on a map (fig. 11). It lies deeper beneath the anticyclonic gyres and shoals beneath the cyclonic gyre. We see the tongues of low salinity (fig. 12) extending along the paths of flow.

This is how the characteristics of the surface water can extend laterally into the deeper waters without vertical overturn to their depths, but by the lateral intrusions we saw on the density section.

The geostrophic flow at 500 decibars (fig. 13) shows how the low-salinity water is carried from its sources around the gyres.

We can pick out an isopycnal that lies along the deeper low-salinity tongues, from the south and look at its depth (fig. 14).

Its shape corresponds closely to the pattern of geostrophic flow. It lies deepest within the anticyclonic gyres and shallowest, or outcrops, within the cyclonic gyres.

In the north we see the low salinity in the northwest, where the isopycnal lies shallow beneath the low salinity surface layers, and extends eastward with the flow (fig. 15).

In the south the isopycnal outcrops at low salinities, and we see the effect of the anticyclonic flow in the South Pacific carrying the low-salinity waters from the Antarctic around the gyre.

There are thus two lateral sources of low salinity in high latitudes, but in between, near the equator, the salinity is high. Where does this come from? This isopycnal does not intersect the sea surface in any area of high salinity. The highest value, between the two sources, is simply a consequence of vertical diffusion from the more saline waters above and below. Both vertical and lateral mixing are required to account for this pattern.

<u>The mid-depth water</u>

We have offered some explanation for these two salinity minima (fig. 6) in terms of air-sea interaction, geostrophic flow, and mixing, both lateral and vertical. But if we look deeper in the Pacific we find a maximum in salinity that does not appear to originate from the sea-surface. It does not fit the simple scheme we have proposed for the upper waters of the Pacific.

As it appears to lie in the circumpolar eastward flow, we look upstream, into the Indian Ocean (fig. 16).

We do not see it originating there.

We see a tongue of low salinity with its origin at the surface in the far south, just as in the Pacific. There is a higher salinity in the north from the Red Sea and Persian Gulf (fig. 17), but it does not extend down to the density of the salinity maximum in the south, and cannot contribute to it (fig. 18 and 19).

The source therefore must be farther upstream, in the Atlantic. So we look farther east still, into the Atlantic, and see a very different picture (fig. 20).

The upper waters of the South Atlantic show a salinity minimum rather like the Pacific and Indian oceans, but the North Atlantic, is entirely different. We see no layer of low-salinity water in the north extending southward, but instead a thicker layer of very warm and saline water. This is clearly the source of the high-salinity water of the deep Pacific and Indian oceans. But why is the North Atlantic so different?

If we look at the surface circulation in the Atlantic (fig. 21) we see that it is also dominated by Trade Winds and Westerlies and it appears to be much the same as in the Pacific, with cyclonic flows in the high latitudes and anticyclonic flows in mid-latitudes. There is no obvious clue from the patterns of surface flow alone as to why the deeper characteristics are so different.

However, (fig. 22) in the far south the density is higher than in the Pacific and in the far north there is a great spill of very dense water from the Greenland Sea, through the Denmark Strait.

The mid-depth waters of the Atlantic Ocean are also very warm as well as saline (fig. 23). Below 1500 m they are the warmest and most saline of the world's oceans.

These high values of heat and salinity are not the result of large-scale and deep convection in the upper layers of the open Atlantic. Instead their sources are the dense waters of the marginal seas—the Mediterranean and Norwegian seas—and the single 1500-2000 m convection within the Labrador Sea.

The Mediterranean, Greenland, and Norwegian seas receive warm and saline surface water from the Atlantic through narrow and shallow passages. Within these seas evaporation and cooling raise their density and they overturn to the bottom.

Those denser waters pour out through the channels beneath the inflowing Atlantic water, at high salinity and temperature, and spread laterally. Their density decreases as they mix with the surrounding water and descend to depths that match their density. Their outflows can be seen on the isopycnal where $\sigma_{1.5}$ is 34.64kg/m^3 (fig. 24 and 25).

The waters within the Mediterranean Sea are the warmest and most saline of the marginal seas of the Atlantic. They flow over the 400 m-deep sill at the Strait of Gibraltar and into the open Atlantic, reaching 1000 m to 2000 m depth as a lateral maximum in salinity (fig. 24 and 25). Part of the outflow extends northward along the eastern boundary. The rest extends westward all across the Atlantic to the western boundary, where it splits, part turning northward and part turning southward far along the boundary to join the circumpolar flow near 50° (fig. 26).

The Labrador Sea overturns to 1000-2000 m in the least saline part of the North Atlantic (fig. 24). It extends eastward near 1600 m along 45°-60°N to about 20°W as both a vertical and lateral salinity minimum (fig. 25). It flows between the high-salinity contributions of the Norwegian and Mediterranean seas.

Dense water from the Norwegian Sea pours out over the ridge east of Iceland, which is less than 500 m deep except for a narrow 1000 m-channel in the east. It can be recognized from 1000 m to as deep as 3000 m by its high salinity as it flows around the ridge extending southwestward from Iceland and turns into the Labrador Sea.

The densest outflows are from the Greenland Sea (fig. 22). They pour southward through the Denmark Strait, between Greenland and Iceland. The sill is only about 500 m in depth, but the density is very high, and the overflow reaches to the bottom south of the ridge. It covers the bottom of the western North Atlantic southward to about 40°N where it meets water extending northward from the Antarctic (fig. 22).

We can see how these sources — the Mediterranean, Labrador, Norwegian, and Greenland seas (fig. 20) provide the thick layer of saline water that extends southward through the South Atlantic and joins the Antarctic Circumpolar flow.

It mixes with the overlying and underlying less saline water and its salinity is reduced during its southward flow, but it is still a vertical and lateral salinity maximum as it joins the circumpolar flow toward the Indian Ocean.

We can follow the high-salinity water by plotting the salinity on an isopycnal, $\sigma 2$=37.0kg/m^3, that extends through the Atlantic, Pacific and Indian oceans (fig. 27 and 28). It lies shallow in the Arctic and Antarctic oceans but deep in between.

The salinity at this density (fig. 28) is highest in the North Atlantic and lowest around Antarctica, but is still recognizable as a lateral maximum in salinity all the way around Antarctica. We can look at the tongue of high salinity along a path from the

North Atlantic, where it is formed, and through the South Atlantic to its junction with the circumpolar flow (fig. 28).

We can see that it extends all the way across the Indian and Pacific oceans and through the Drake passage (fig. 29).

Here (fig. 30) the deeper fields of salinity and temperature are plotted against density to avoid the ups and downs of the isopycnal along its path.

We see the highest values from the Mediterranean and Norwegian sea outflow. The maximum value of salinity decreases along the flow, especially in the South Atlantic, beneath the Intermediate Water.

The vertical maximum finally disappears as it flows through the Drake Passage into the more saline Atlantic where it had originated.

Water of this density, of course continues to flow eastward with the circumpolar current, but part of it turns northward into the South Atlantic east of the southward flow along the western boundary, and into the North Atlantic (fig. 28). It has lost so much salt in its trip around the world that it can be recognized now, in the high salinity of the Atlantic, as a lateral salinity minimum east of the southward flow. It extends northward beyond 20°N towards the origin of the salinity maximum. (The ocean circulates!) This emphasizes that not all of the deeper waters in the Atlantic are moving southward to the Antarctic, with northward flow only at abyssal depths. This is seen even more clearly along a slightly denser isopycnal (fig. 31).

The deepest flow

We have accounted for the presence of the subsurface salinity minima of the southern hemisphere and the North Pacific, and the great salinity maximum from the North Atlantic. But we have not dealt with the densest and deepest waters.

Most of the surface waters around Antarctica are very cold, but much too low in salinity to become dense enough, even when cooled to freezing, to reach the density found at great depths. Some source other than local processes must add the extra salt that raises the density to the values seen at abyssal depths.

The densest and coldest of the bottom waters of the open ocean are found in the three great basins extending around Antarctica; the Weddell-Enderby, the Australian-Antarctic, and the Southeast Pacific basins (fig. 32). (Fig. 32 and 33 were prepared before the change in the equation of state. Modern values of potential density would be lower by .046 kg/m^3 and potential temperature higher by .032° C than the values shown here.)

They are not formed by top-to-bottom convection in the open ocean, but along the shelves and slopes at various places around the Antarctic Continent.

They are mixtures of the very cold and low-salinity surface waters around the continent with the warmer and more saline waters of the circumpolar current, which carry the warm and saline water from the North Atlantic.

These mixtures, cooled further at the sea surface, and with some brine released by freezing at the surface along the shelves of Antarctica, become dense enough to flow down the slopes and into the three great basins.

The Weddell Sea produces the densest waters (fig. 22). Some of the circumpolar flow joins the Weddell Sea Gyre and turns southward and then westward toward the southwest Weddell Sea. Its flow along the coast of Antarctica carries water that is more saline than can be produced locally. We see that the isolated high-salinity feature in the Weddell Sea (fig. 20) is not separate, but a westward turnback of the saline circumpolar flow (fig. 28 and 30).

There is also such a turnback in the Ross Sea, which leads to the dense bottom waters found there, and there are other sites of formation along the continental shelf of the Australian-Antarctic Basin.

All of these basins reach depths of more than 4000 m but the ridges separate them from each other and limit the exchange of the densest waters with the northern waters (fig. 32). Only after mixing with the overlying water has lowered their density can they flow over the various ridges and rises into the deep open ocean.

They spread northward through the deepest passes between the various ocean ridges and become warmer and less dense by mixing with the overlying water. Only in the far north Atlantic, where some dense waters from the Greenland Sea pour down to the bottom, do we find another dense source, whose signal is mixed away as it passes southward, and is lost near 40°N.

In the Atlantic and Pacific the densest northward flow is from the Antarctic along the western boundary, but the Indian Ocean receives its densest and coldest waters from both the Weddell-Enderby Basin along the western boundary and the Southeast Indian Basin in the east (fig. 32 and 33).

The abyssal density is lowest and the temperature highest in the basins east of the Mid-Atlantic Ridge and the East Pacific Rise, whose openings are not deep enough to admit the denser water.

The abyssal salinity (fig. 34) is also low around Antarctica from the mixture of cold surface water with the warm and saline North Atlantic waters. It increases northward everywhere except in the North Pacific, beyond the extent of the great layer of maximum salinity.

It is not only the dense water from the bottom of the Antarctic that extends northward into the three oceans. All of the water beneath the Weddell Sea and Ross Sea gyres, even near the surface, is dense enough to extend laterally to more than 4000 m depth in the northern oceans.

Along its northward flow it mixes with the overlying warmer and less dense water and finally becomes low enough in density to rise above the northward flow and return southward toward the Antarctic. This southward flow is not a single stream but wends its way past and around the interlocking gyres and becomes part of the wind-driven flow.

Along these paths the cyclonic gyres draw water to or near the sea surface, as does upwelling along the equator, the west-wind drifts, and the eastern boundaries of mid-latitudes.

Summary

The large-scale circulation we observe is the sum of three accelerations: the winds, the exchange of heat and water through the sea surface, and the rotation of the earth.

The Trade Winds and Westerlies cause the great eastward and westward flows across the ocean. These flows are turned back at the continental boundaries, forming anticyclonic gyres in mid-latitudes and cyclonic gyres in high latitudes.

The earth's rotation adds a cum sole acceleration of the flow that is toward the axes of the anticyclonic gyres and away from the axes of the cyclonic gyres. This raises the anticyclonic and lowers the cyclonic gyres.

The ocean is stratified in temperature, salinity, and density. The exchange of heat and water makes the surface water in higher latitudes denser than in lower latitudes. Denser waters from the far north and south spread equatorward beneath the less-dense waters of the tropics, into the subtropics.

We have recognized several layers of minimum salinity that flow beneath the surface. They extend equatorward from the high latitude cyclonic gyres of the North and South Pacific, the South Atlantic, and the Indian oceans to more than 1000 m.

Beneath these layers there is a layer of maximum salinity that is formed in the North Atlantic from the waters of the marginal seas—the Mediterranean, Labrador, and Greenland and Norwegian seas, and that can be seen flowing southward in the Atlantic, joining the circumpolar flow all around Antarctica. Some of is extends to the shelves and slopes of the continent where they are cooled further, and with the addition of brine, they form the densest waters of the open ocean. These sink to the bottom and flow northward along the deepest paths and cross the equator and into the northern oceans.

Acknowledgements

This essay is taken from the final lecture in an introductory course I gave for many years. It is not based upon my work alone but upon the work of many other investigators as well. As a lecture it did not include citations.

Illustrations and discussions of the non-conservative characteristics (oxygen, nutrients, isotopes) were included in the original lecture, but for brevity they are omitted here. However, they were included in the three publications on the separate oceans listed below, which amplify the text and acknowledge the contributions of other investigators.

Reid, Joseph L. 1994. On the total geostrophic circulation of the North Atlantic Ocean: Flow patterns, tracers, and transports. Prog. Oceanogr., 33, 1-92.

Reid, Joseph L. 1997. On the total geostrophic circulation of the Pacific Ocean: Flow patterns, tracers, and transports. Prog. Oceanogr., 39, 263-352.

Reid, Joseph L. 2003. On the total geostrophic circulation of the Indian Ocean: Flow patterns, tracers, and transports. Prog. Oceanogr., 56, 137-186.

This work was supported by the National Science Foundation and the Integrative Oceanographic Division of the Scripps Institution of Oceanography. I wish to acknowledge the assistance given by Arnold Mantyla in selecting the data and by David Newton for writing the various programs. I wish to acknowledge especially Sarilee Anderson for the great skill in handling the various data formats, in arranging the data and calculating and plotting the data points and in preparing the figures.

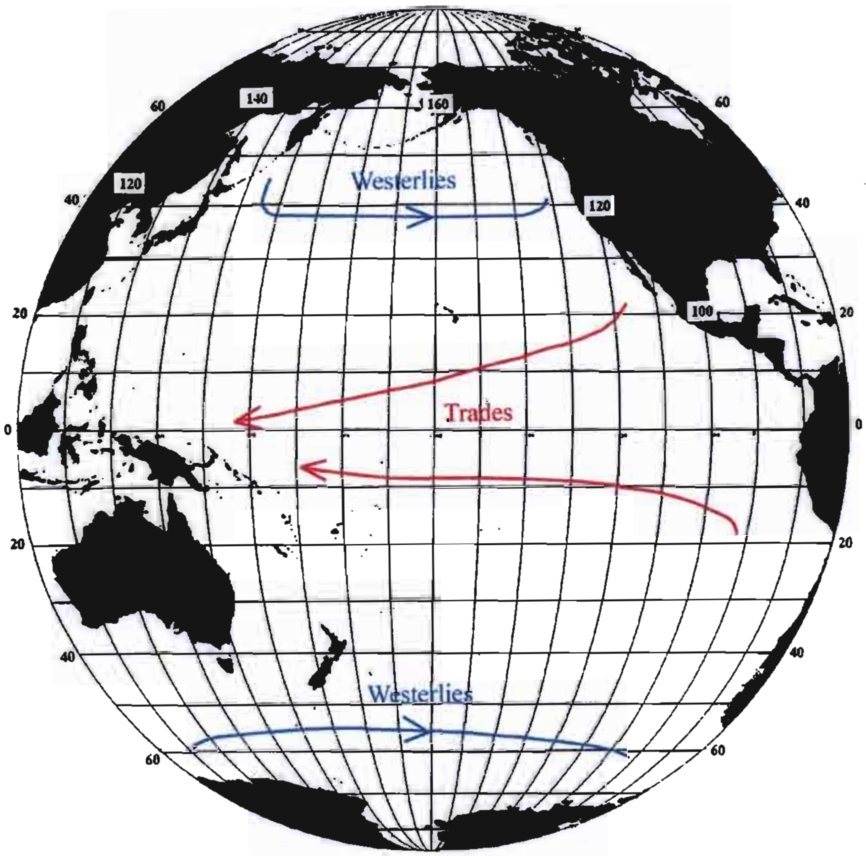

Fig. 1 Pacific Ocean winds

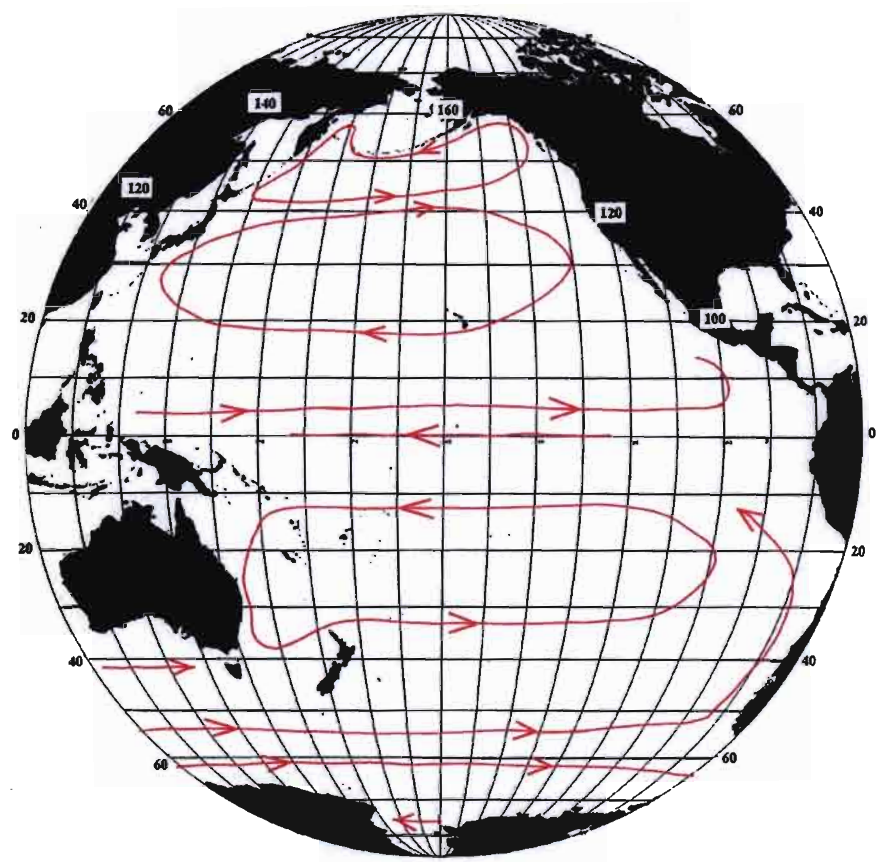

Fig. 2 Pacific Ocean circulation

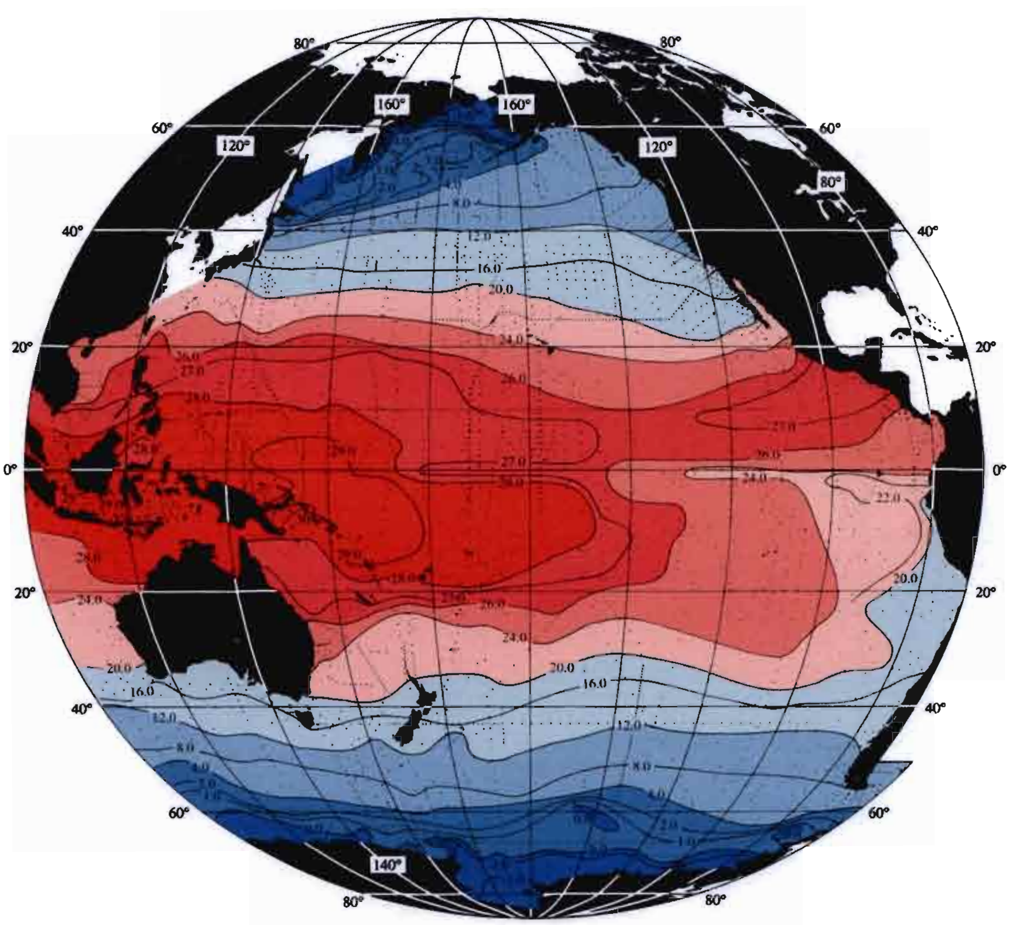

Fig. 3 Pacific Ocean sea surface temperature in winter

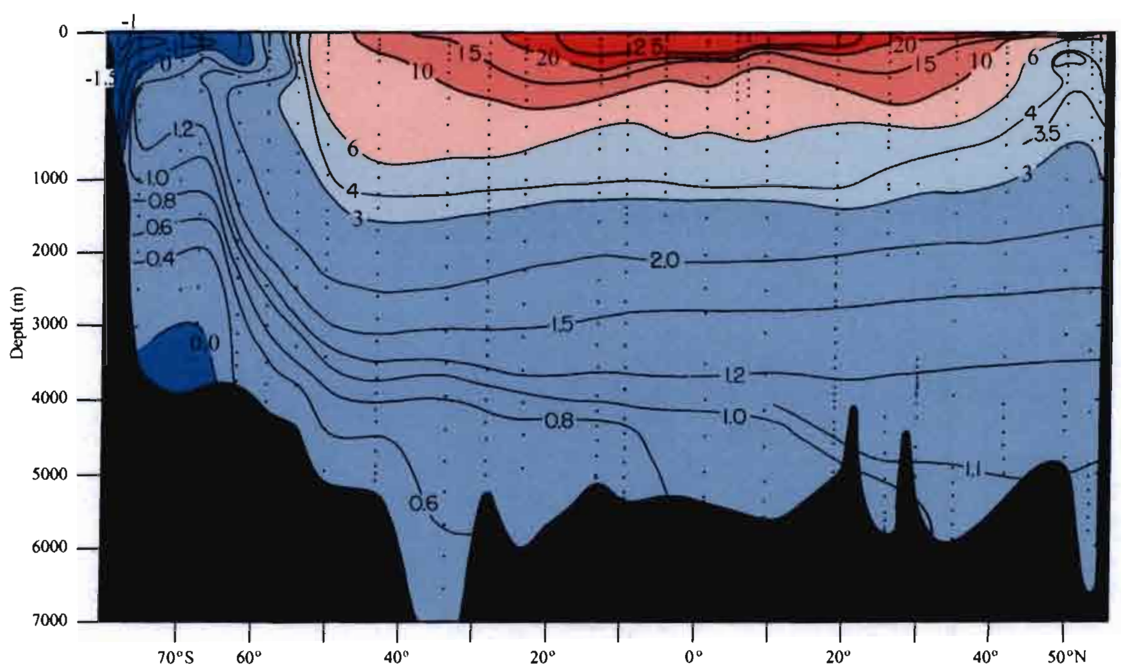

Fig. 4 Pacific Ocean potential temperature along 170°W

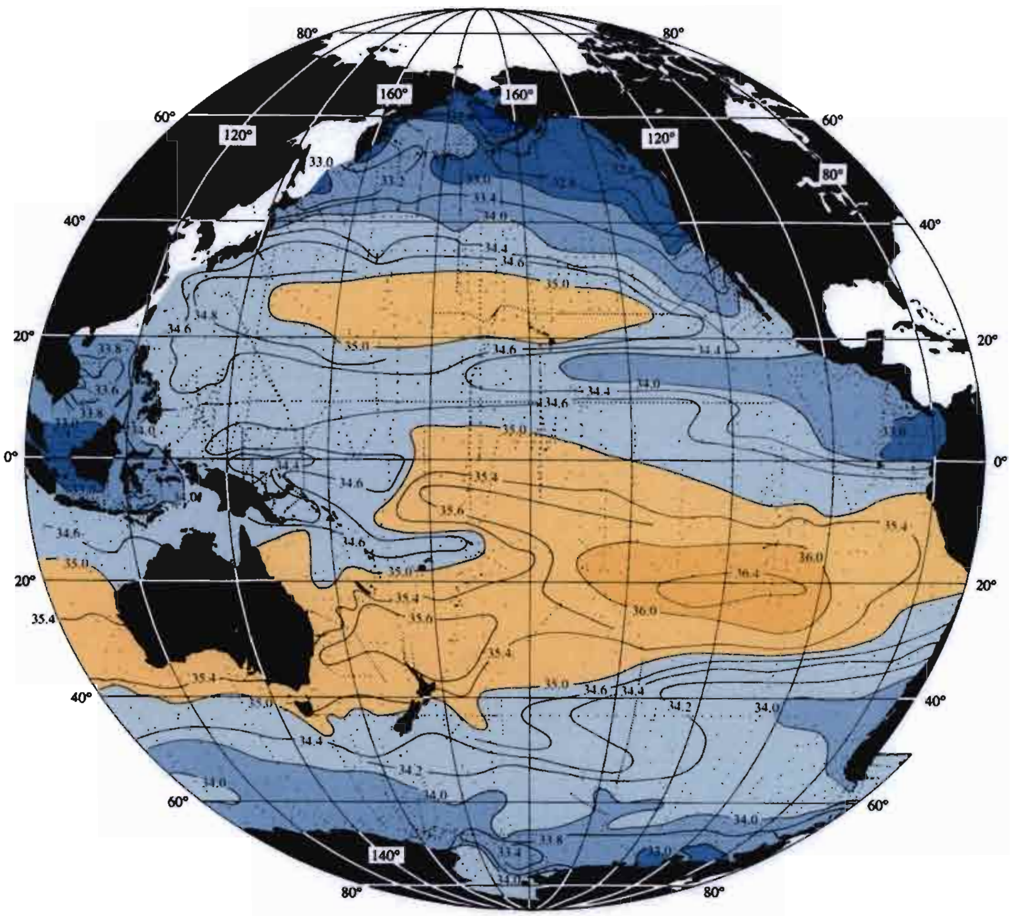

Fig. 5 Pacific Ocean sea surface salinity in winter

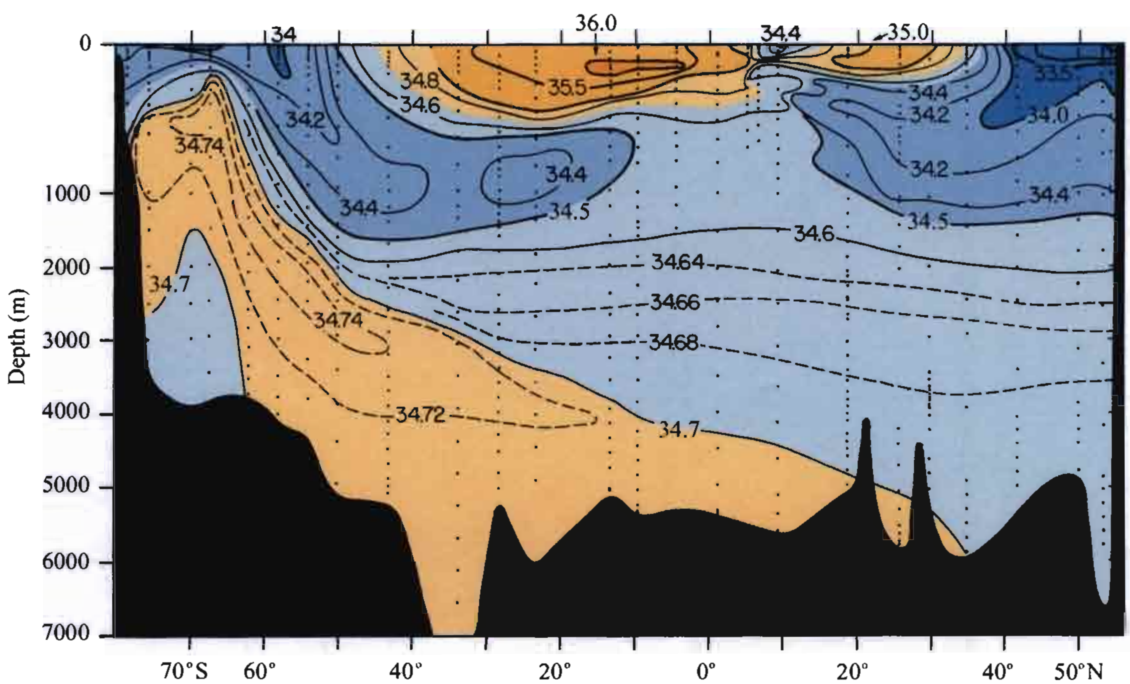

Fig. 6 Pacific Oc ean salinity along 170°W

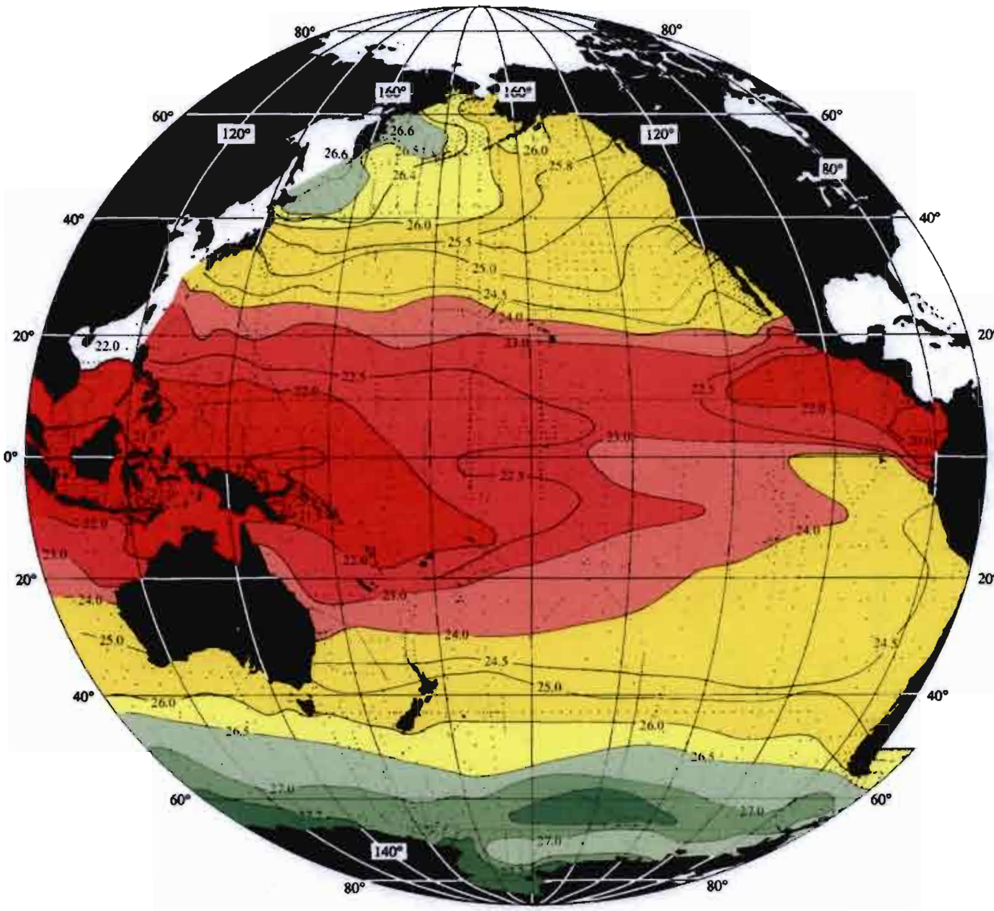

Fig. 7 Pacific Ocean sea surface density anomaly in winter

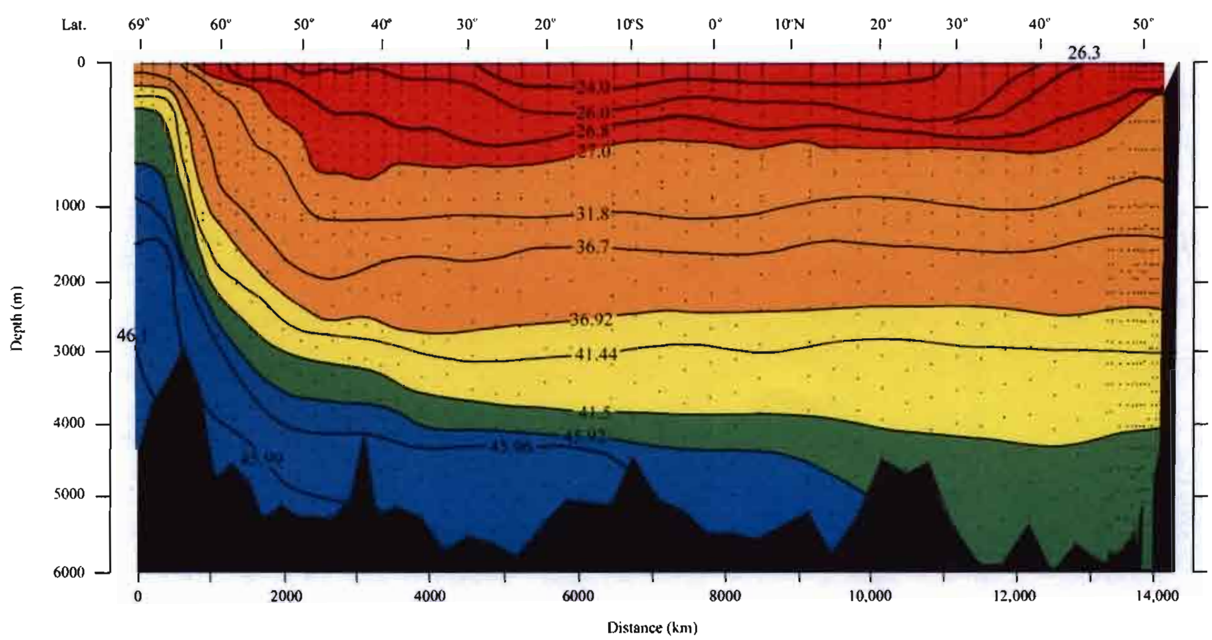

Fig. 8 Pcific Ocean potential density anomaly along 170°W

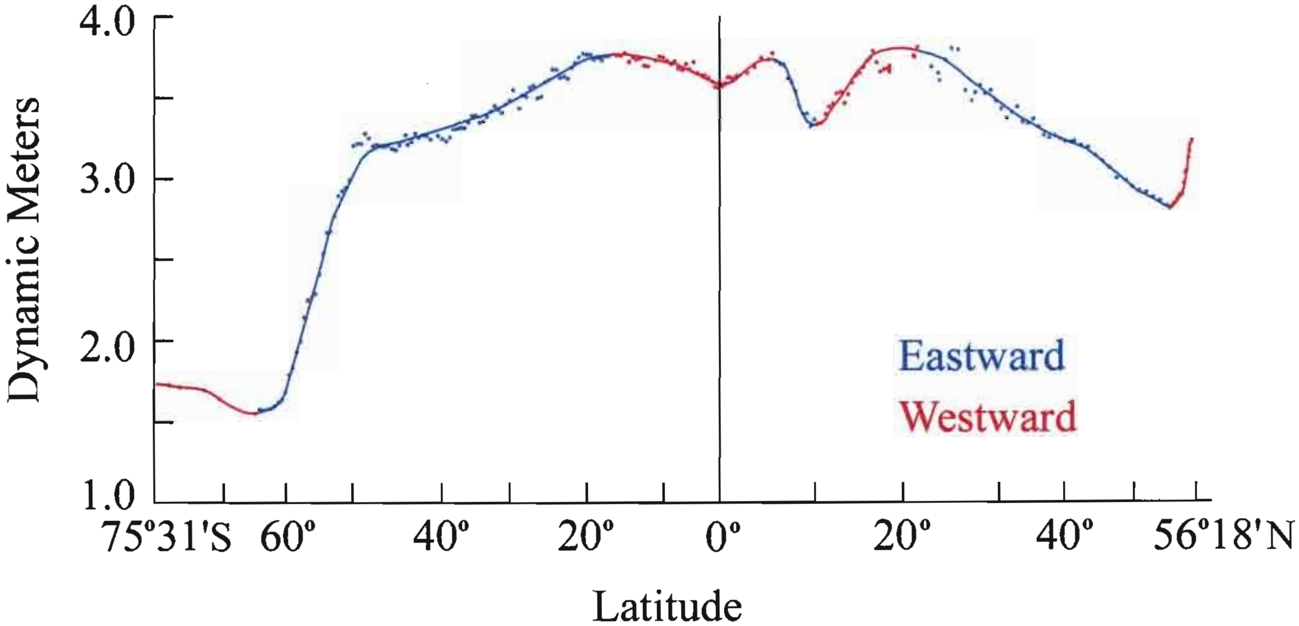

Fig. 9 Steric height, dynamic meters, of the sea surface along 170°W

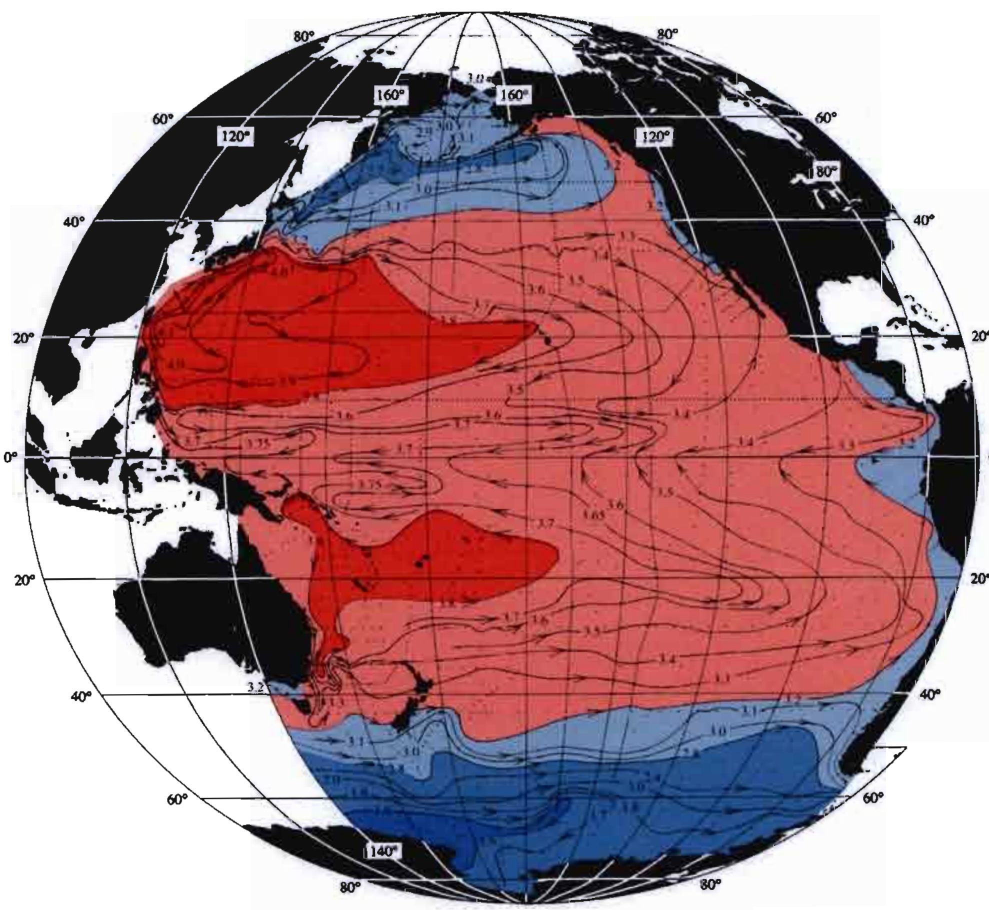

Fig. 10 Pacific Ocean adjusted steric height, dynamic meters, of the sea surface

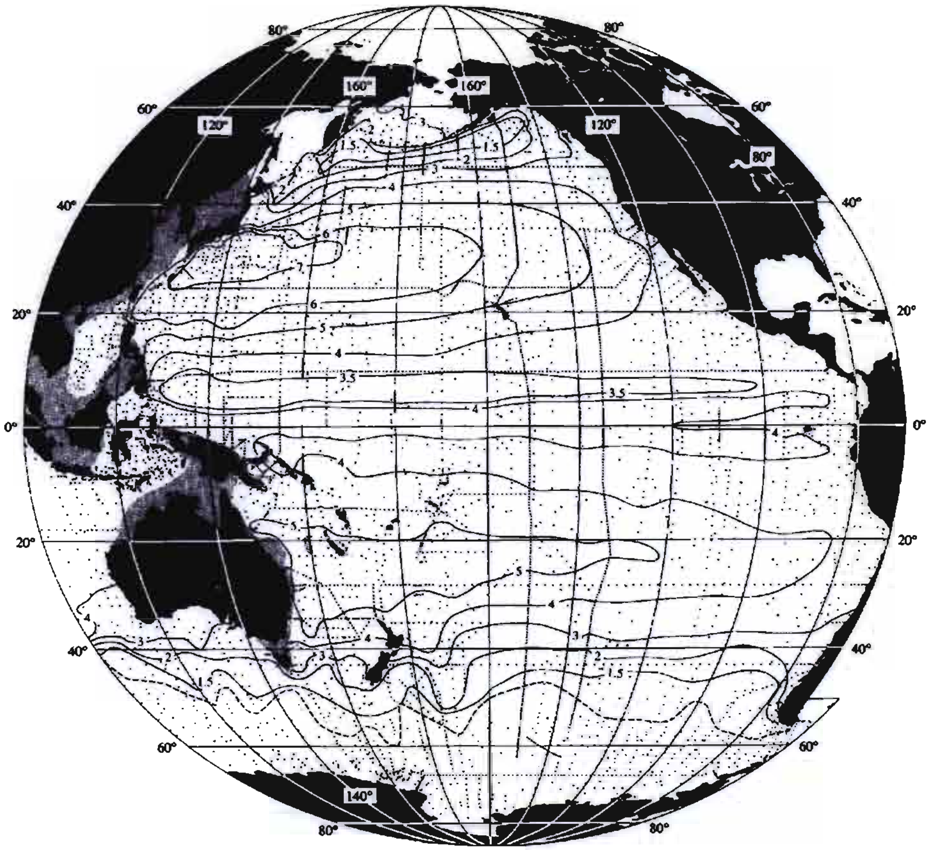

Fig. 11 Pacific Ocean depth (hm) where σ_0 is 26.80 kg/m^3

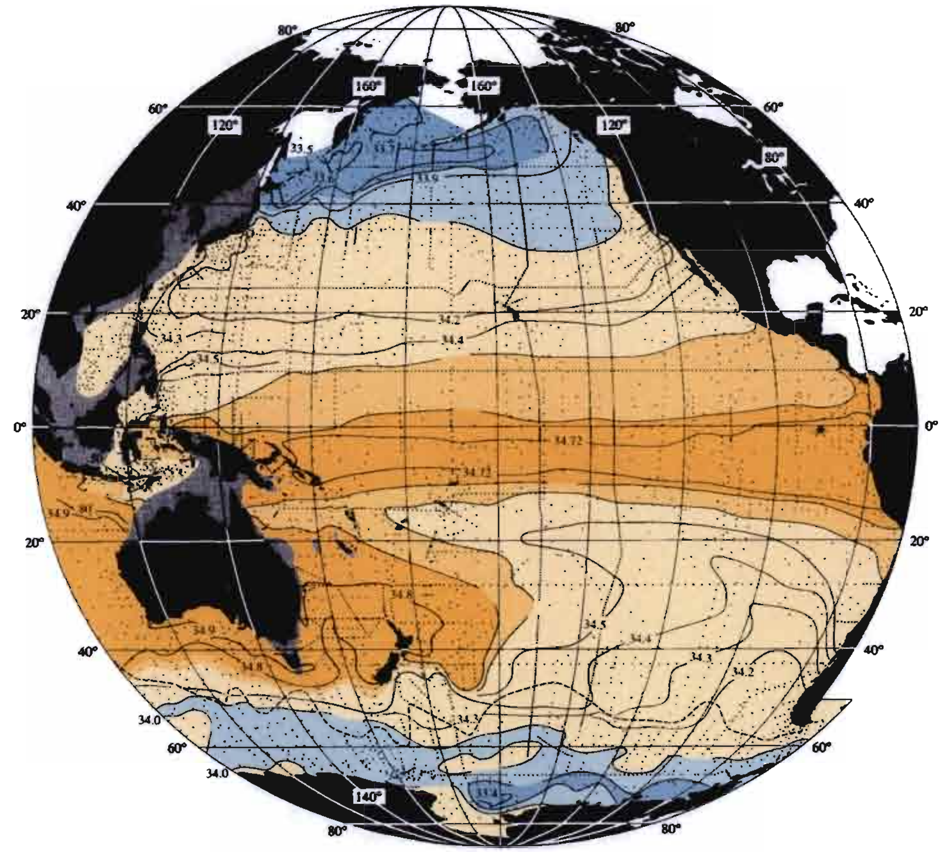

Fig. 12 Pacific Ocean salinity where σ_0 is 26.80 kg/m^3

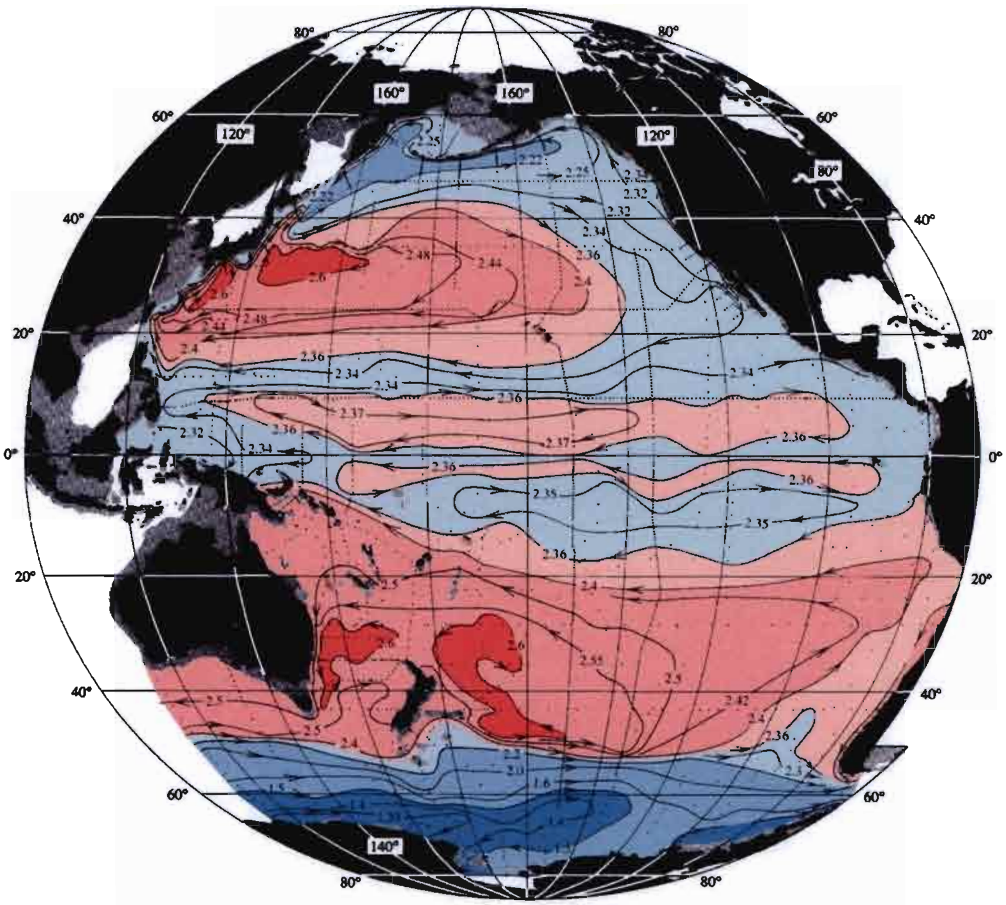

Fig. 13 Pacific Ocean adjusted steric height at 500 decibars

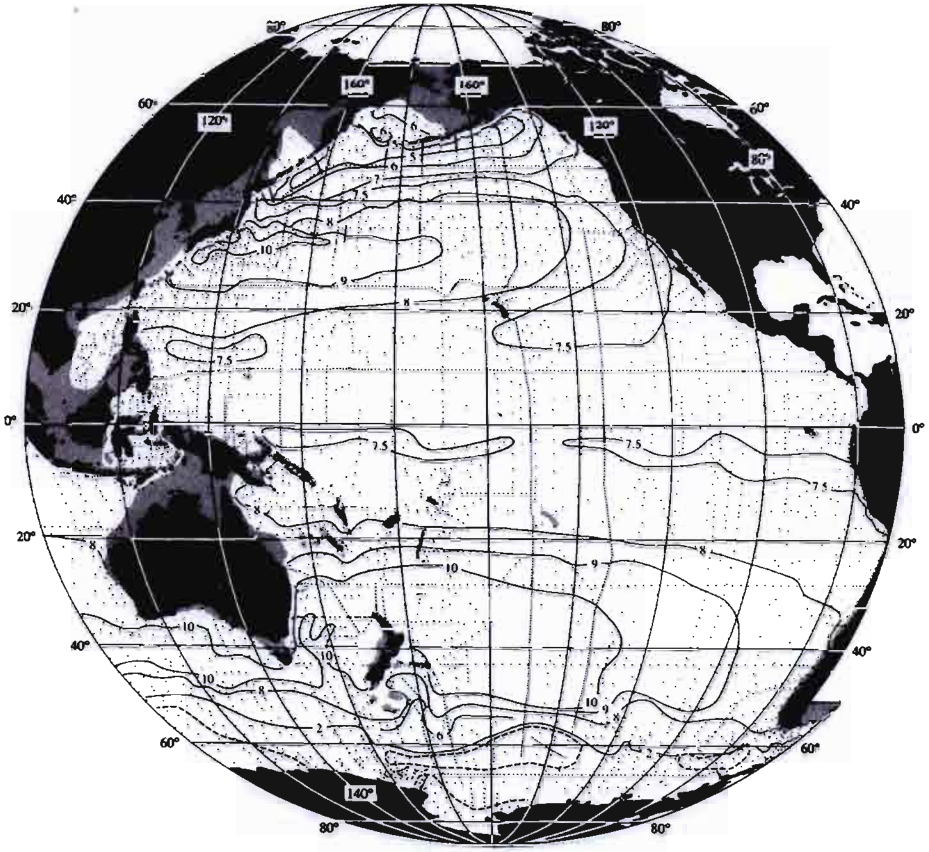

Fig. 14 Pacific Ocean depth (hm) where σ_1 is 31.80 kg/m^3

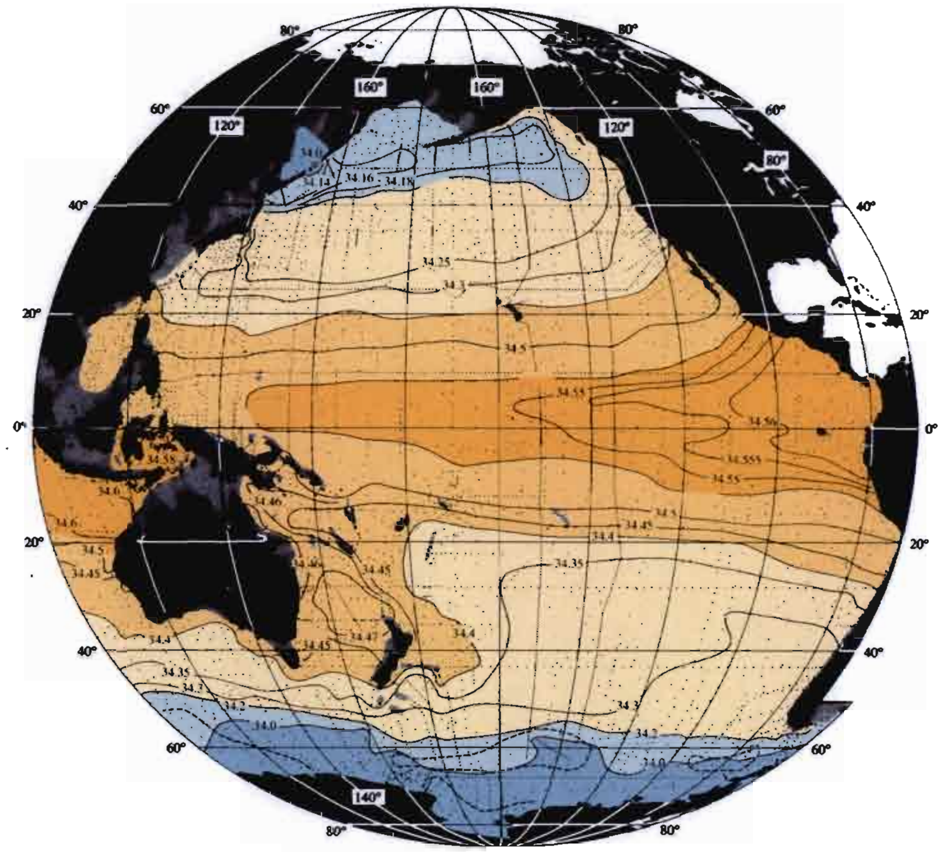

Fig. 15 Pacific Ocean salinity where σ_1 is 31.80 kg/m^3

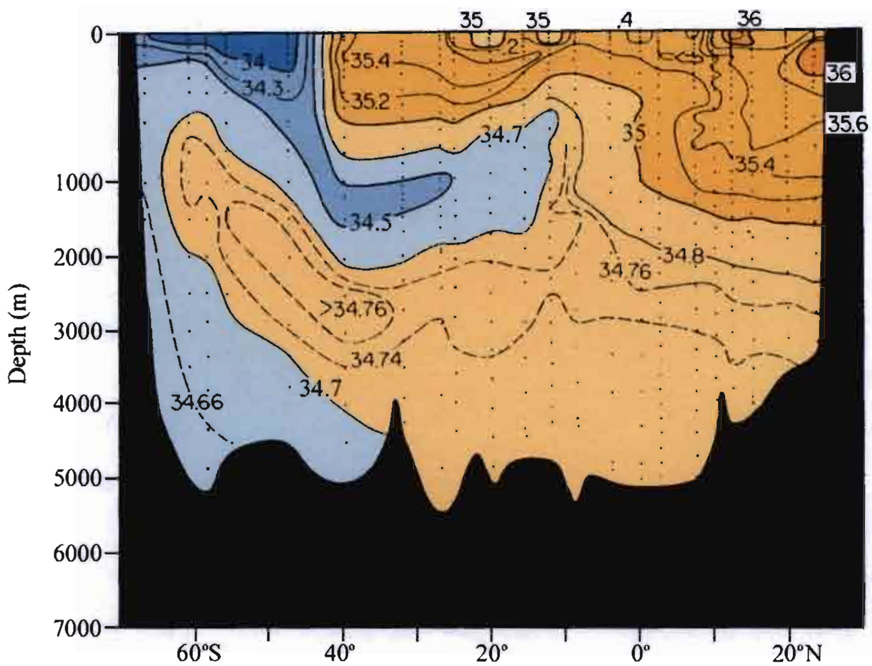

Fig. 16 Indian Ocean salinity along north-south section

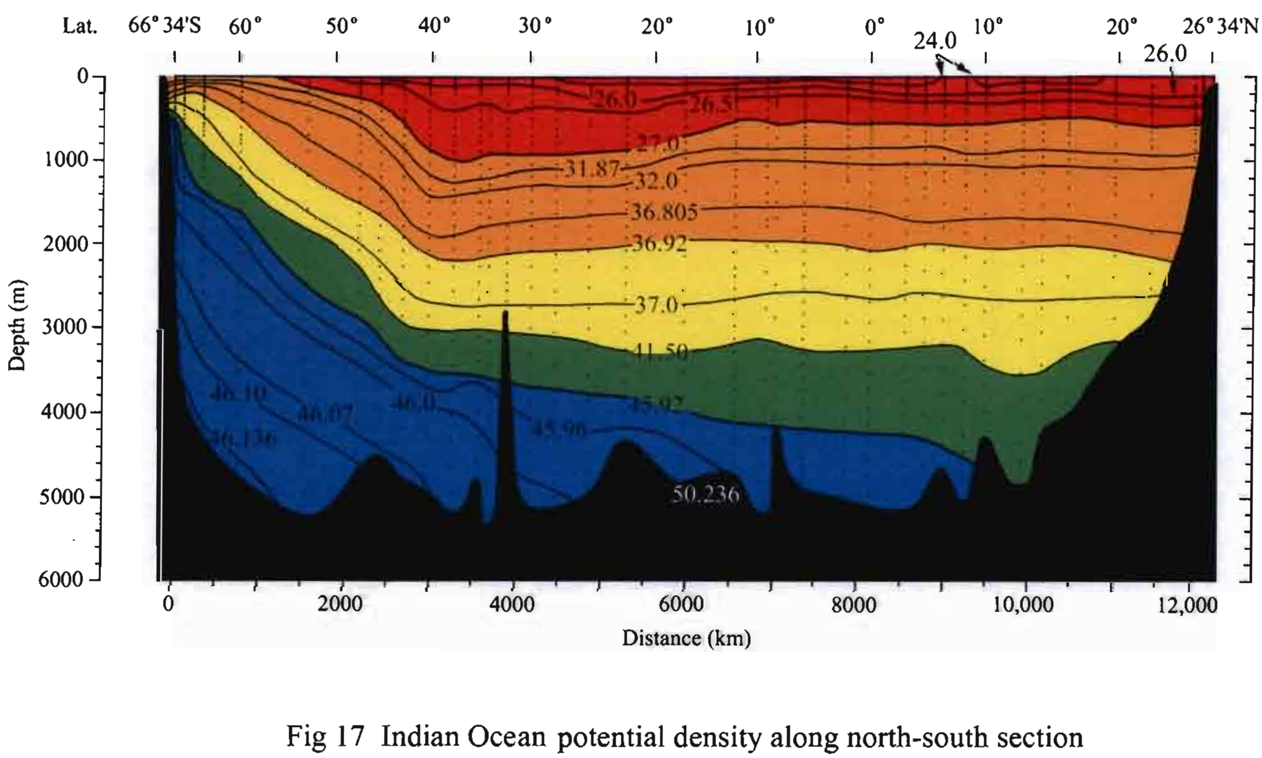

Fig 17 Indian Ocean potential density along north-south section

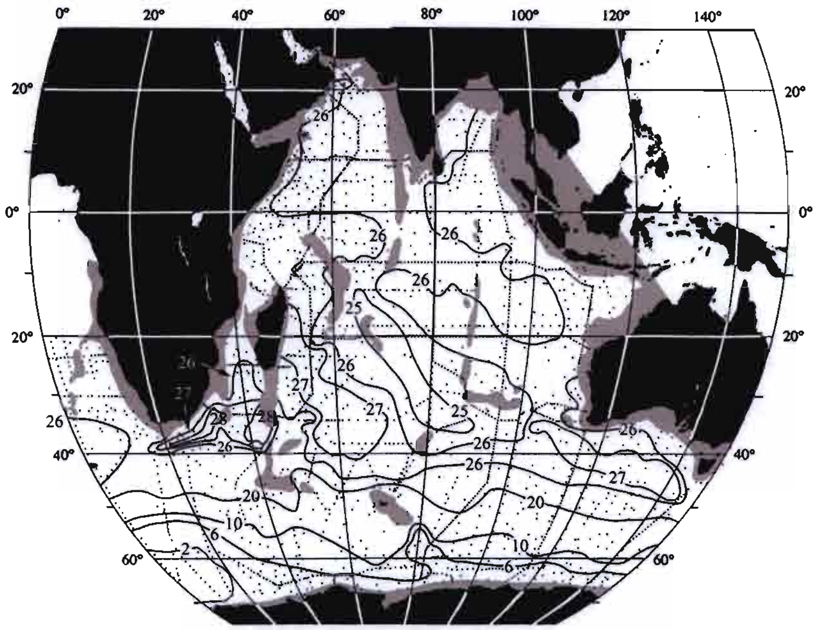

Fig. 18 Indian Ocean depth (hm) where σ_2 is 37.0 kg/m^3

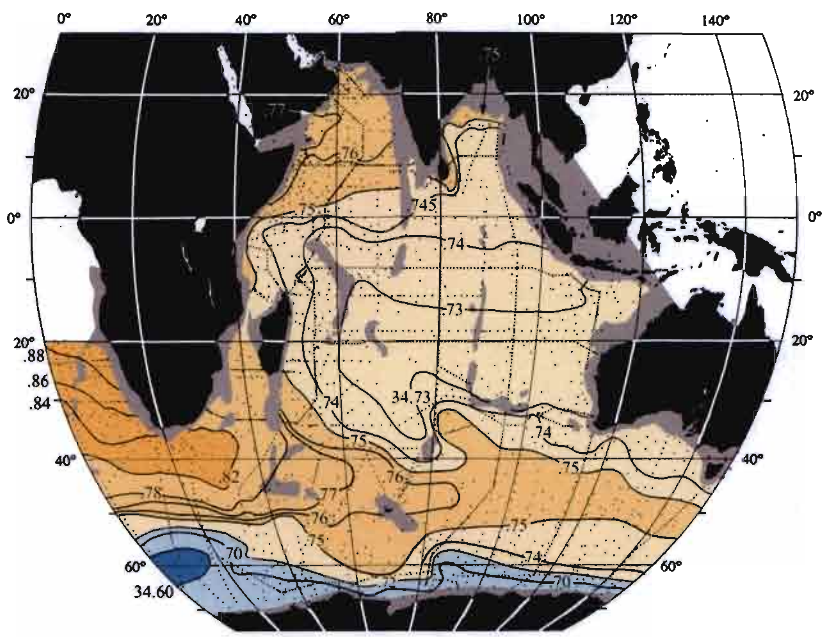

Fig 19 Indian Ocean salinity where σ_2 is 37.0 kg/m^3

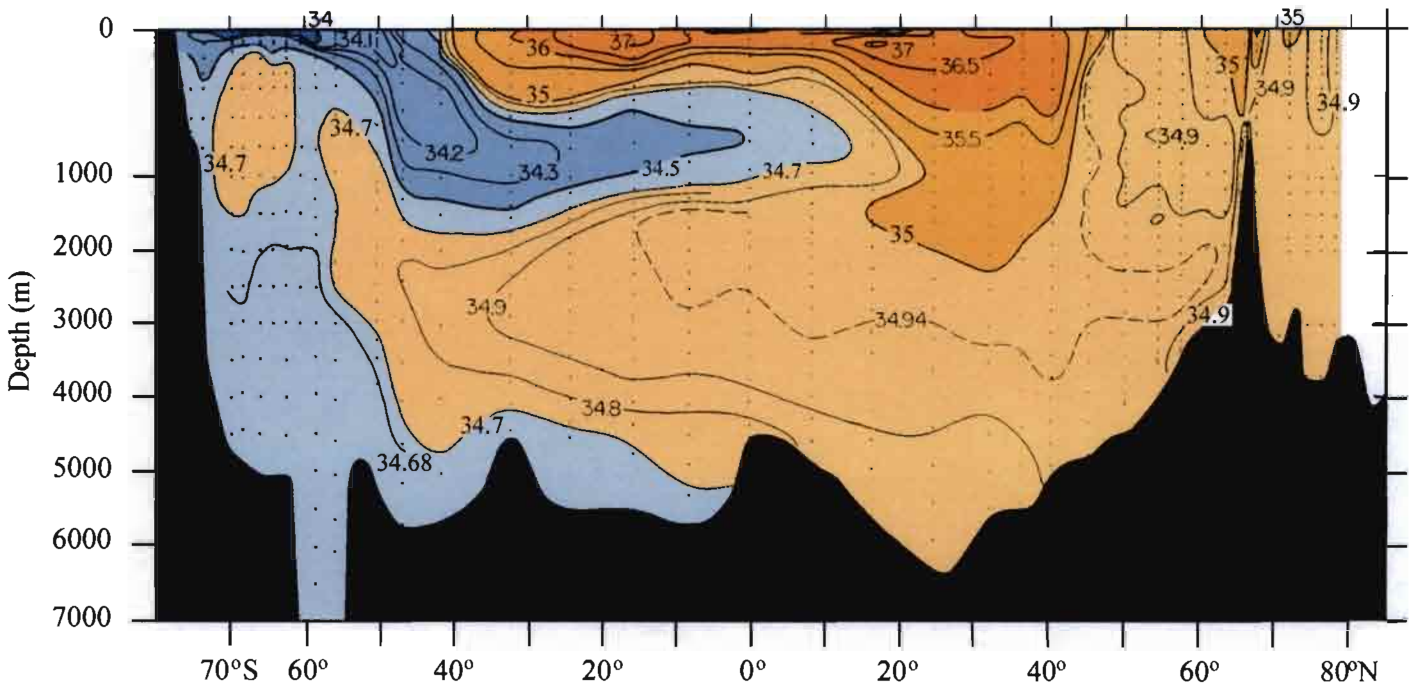

Fig. 20 Atlantic Ocean salinity along north-south section, west of Mid-Atlantic Ridge

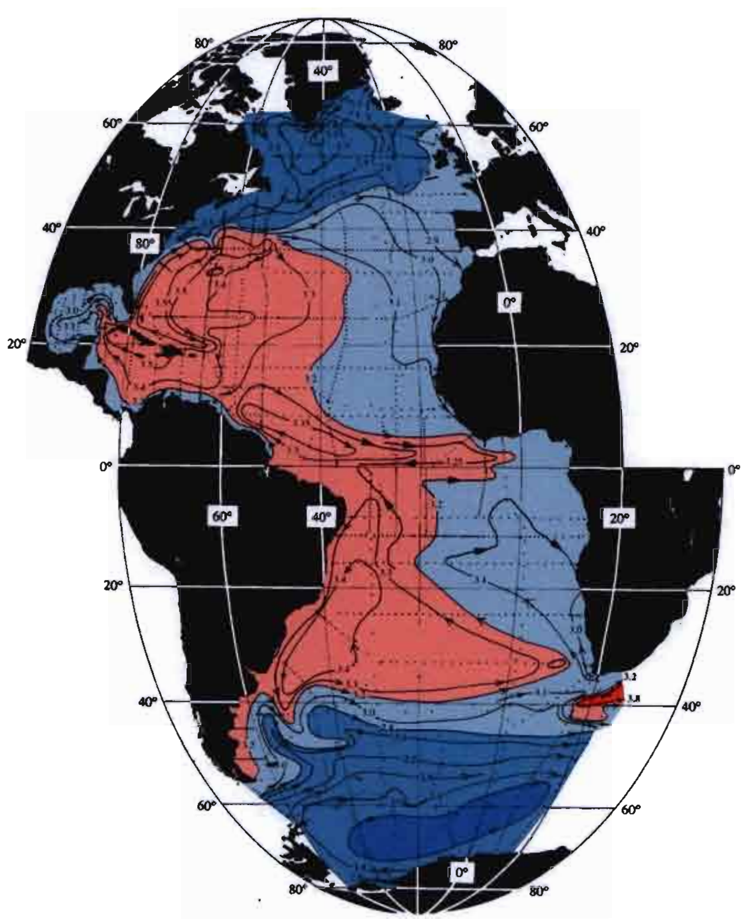

Fig. 21 Atlantic Ocean adjusted steric height at the surface

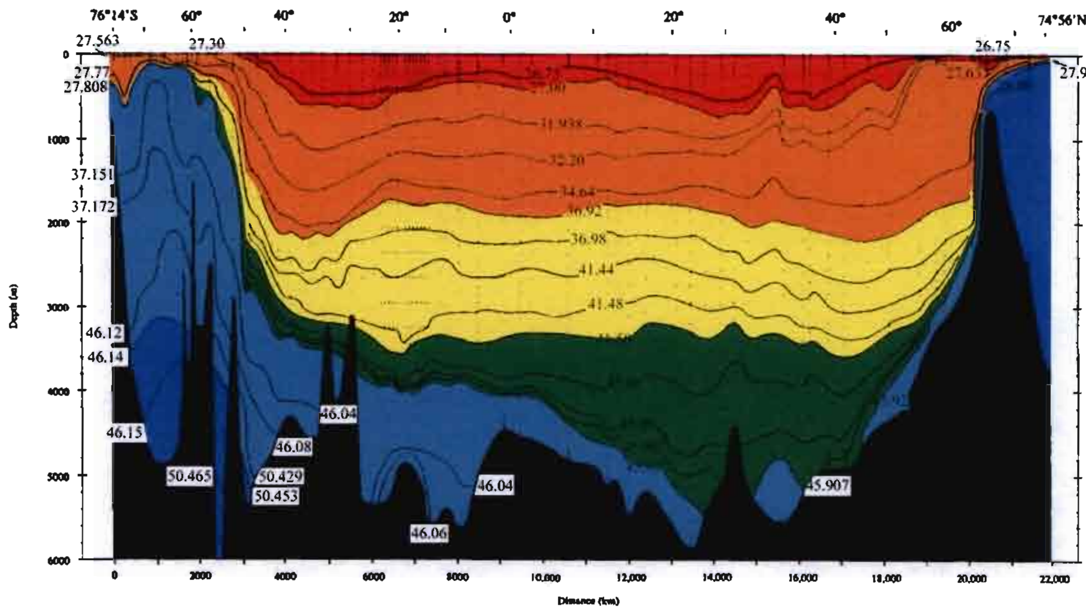

Fig. 22 Atlantic Ocean potential density along north-south section west of Mid-Atlantic Ridge

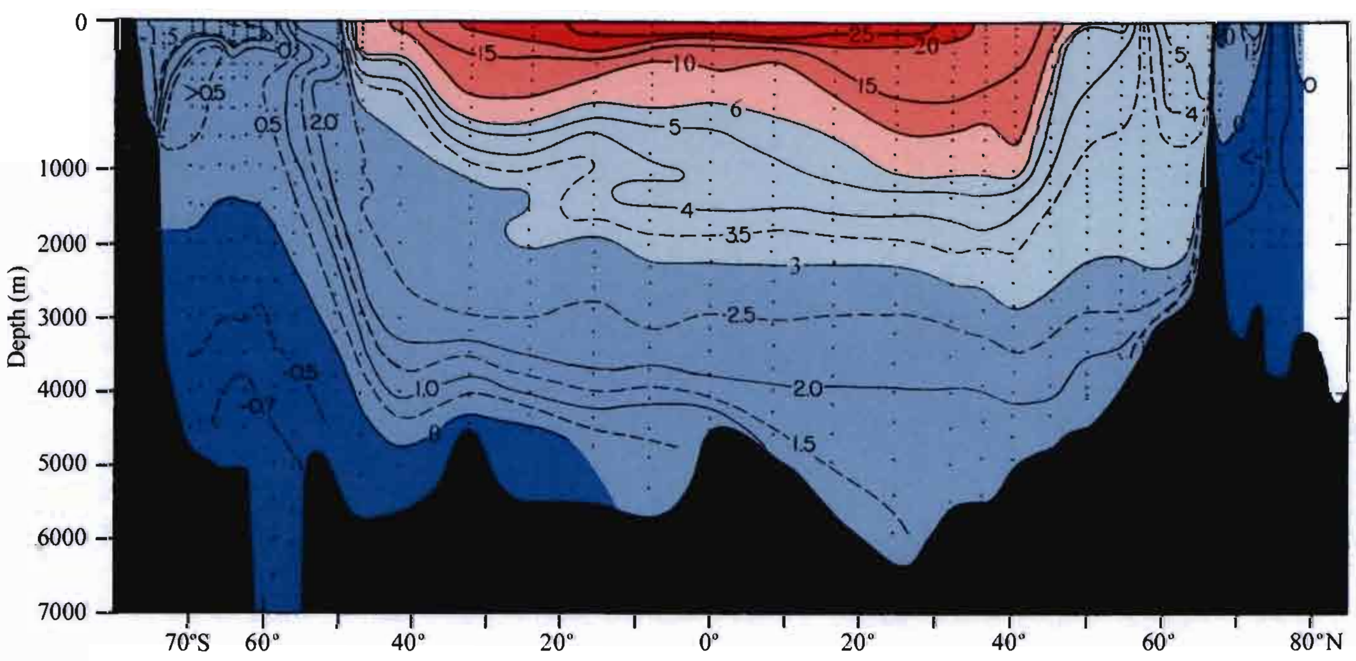

Fig. 23 Atlantic Ocean potential temperature along north-south section west of Mid-Atlantic Ridge

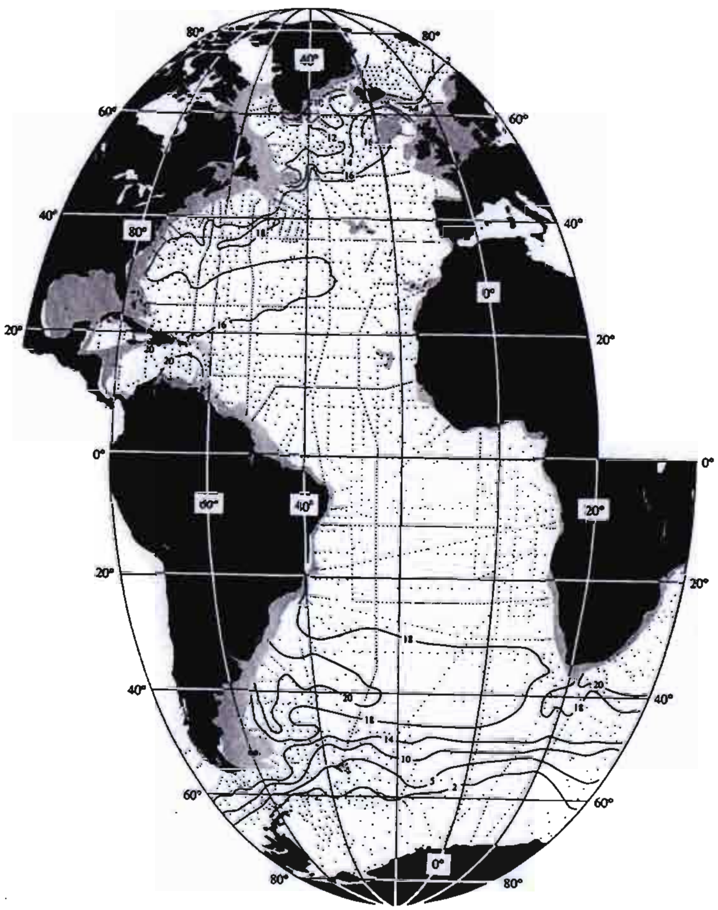

Fig. 24 Atlantic Ocean depth (hm) where $\sigma_{1.5}$ is 34.64 kg/m^3

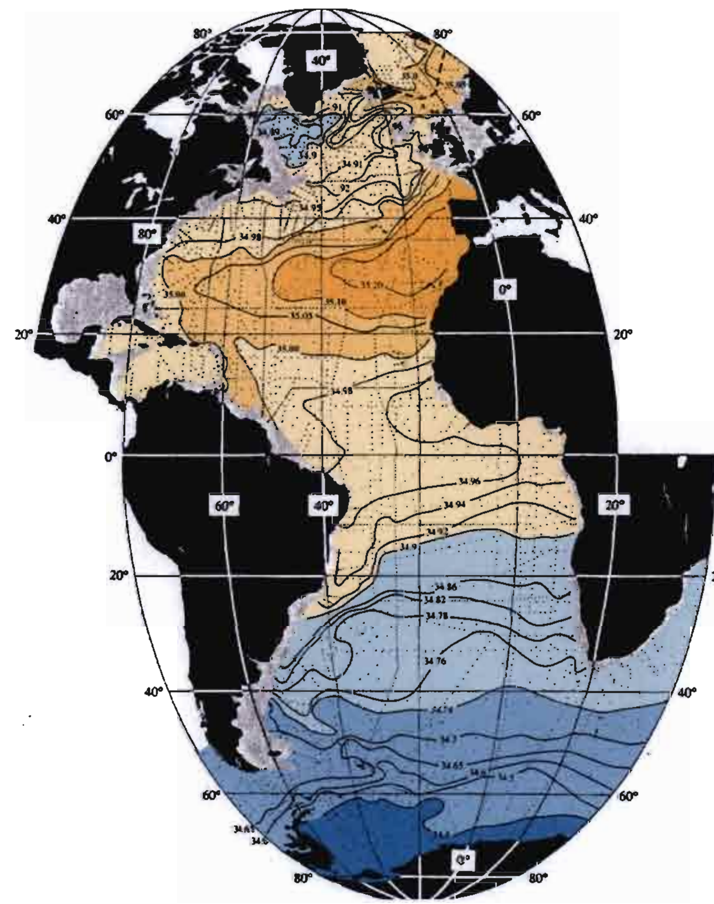

Fig. 25 Atlantic Ocean salinity where $\sigma_{1.5}$ is 34.64 kg/m^3

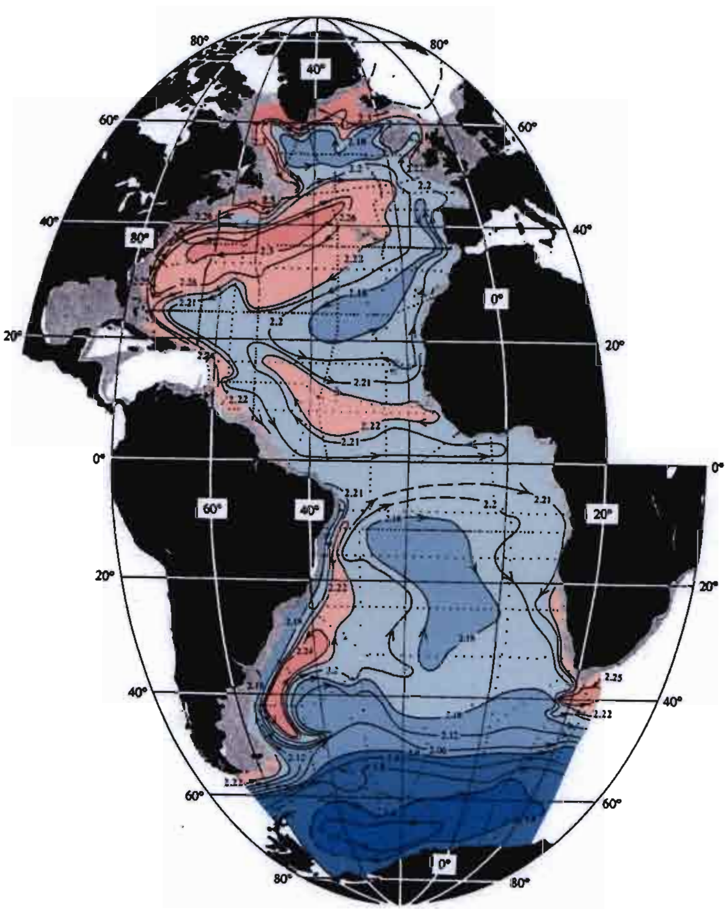

Fig. 26 Atlantic Ocean adjusted steric height where $\sigma_{1.5}$ is 34.64 kg/m^3

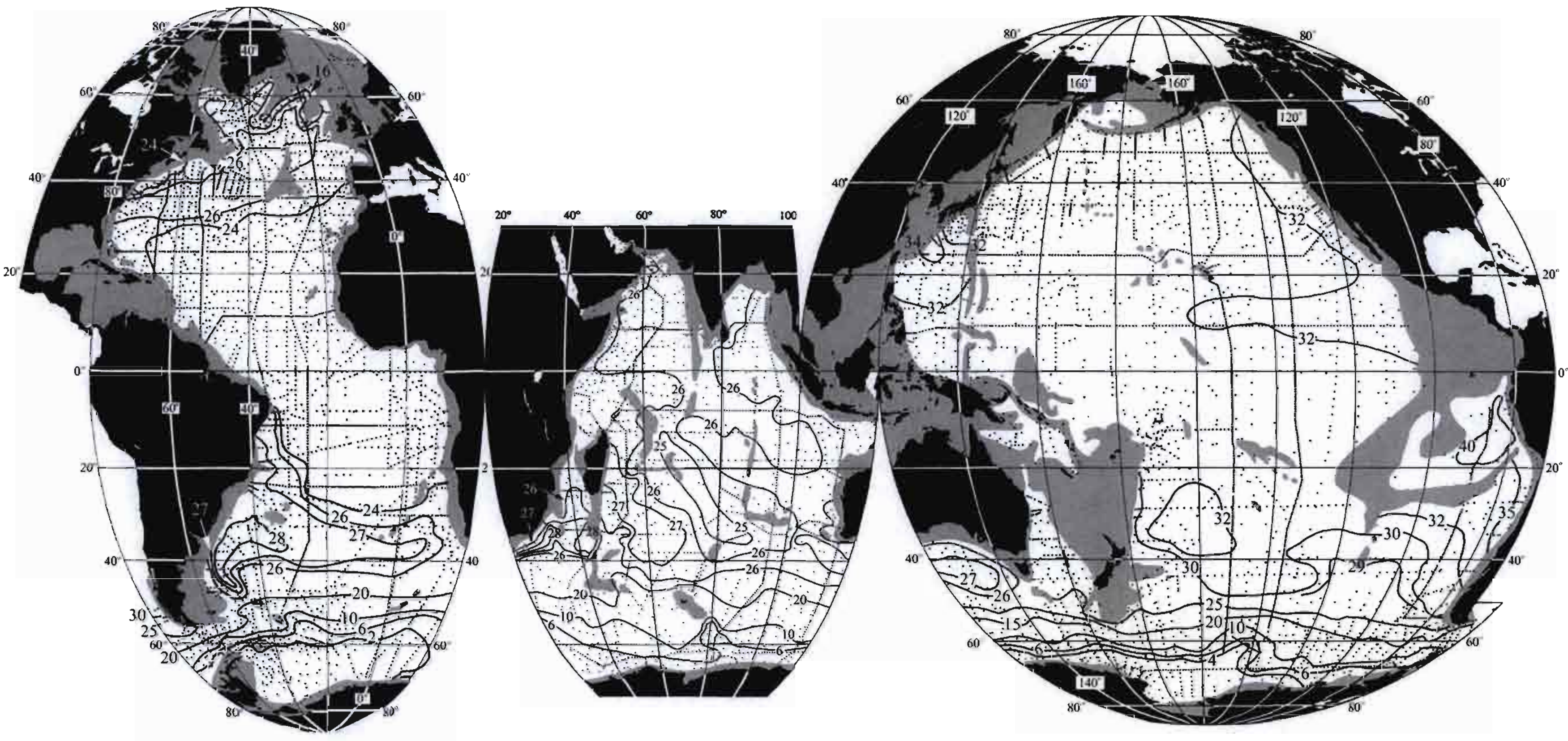

Fig. 27 World Ocean depth (hm) where σ_2 is 37.00 kg/m^3

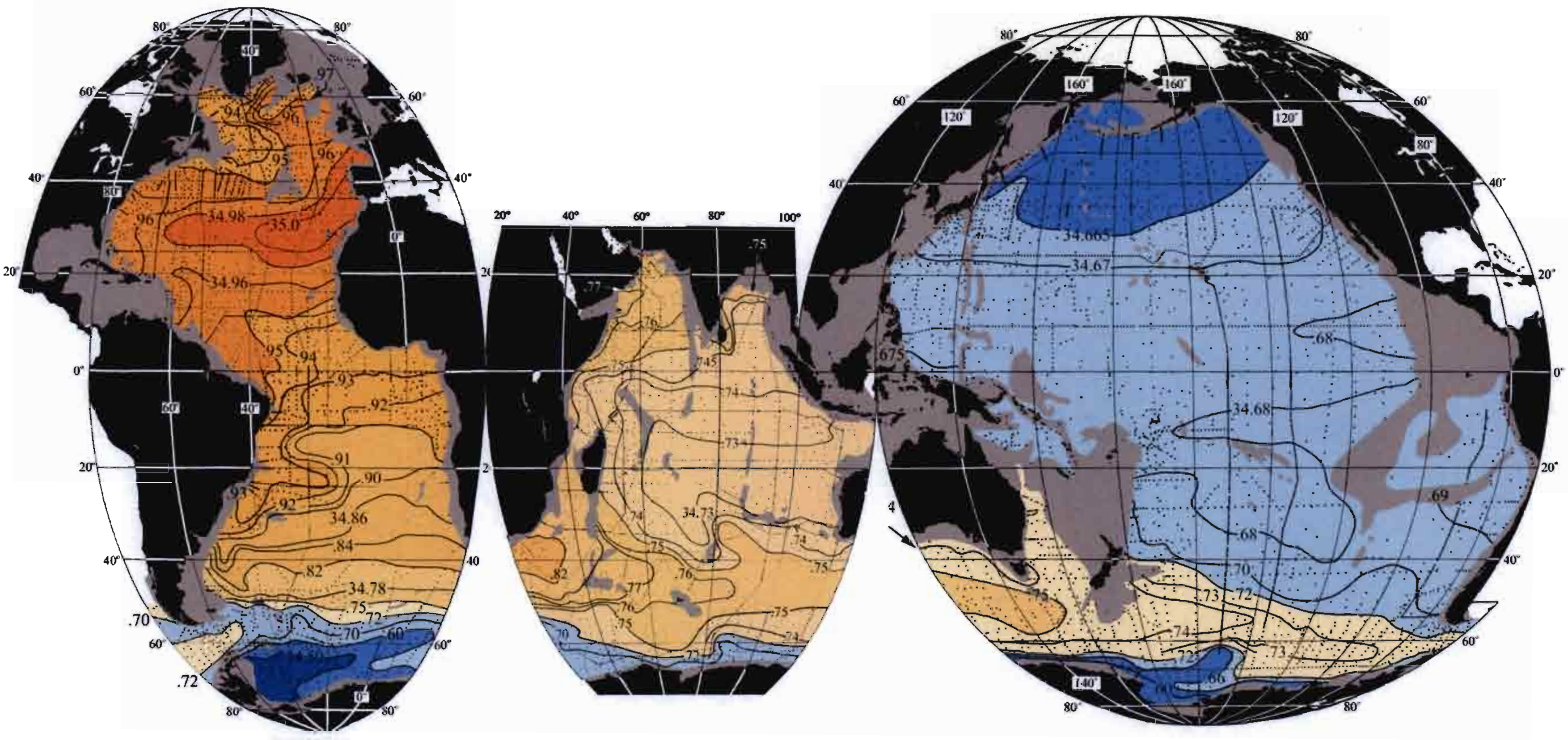

Fig. 28 World Ocean salinity where σ_2 is 37.00 kg/m^3

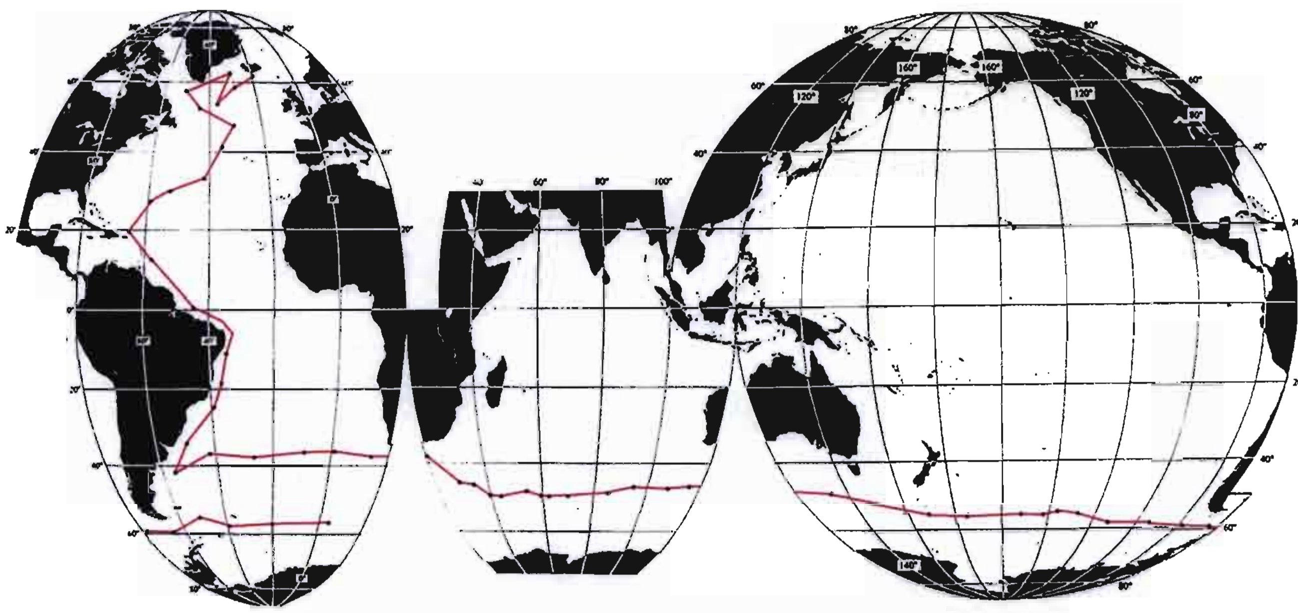

Fig. 29 Section from North Atlantic through Pacific

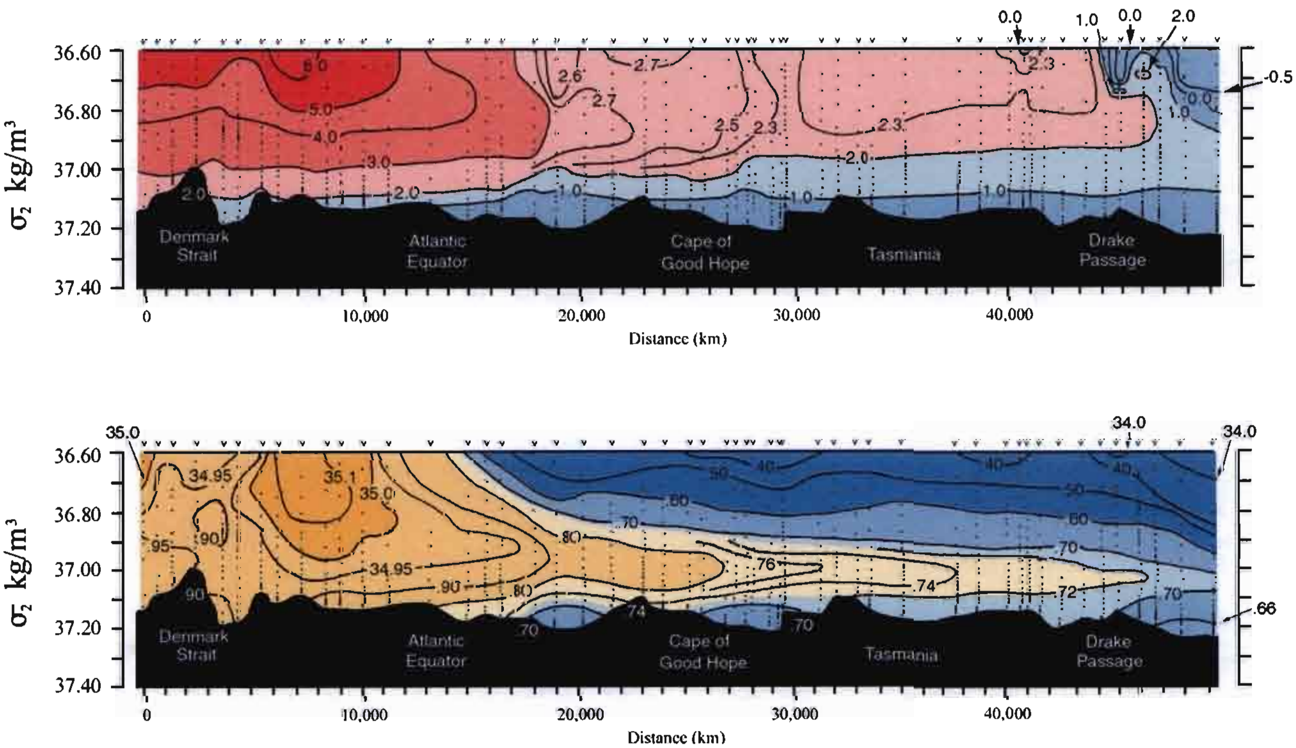

Fig. 30 Potential temperature and salinity along the path, with σ as the ordinate

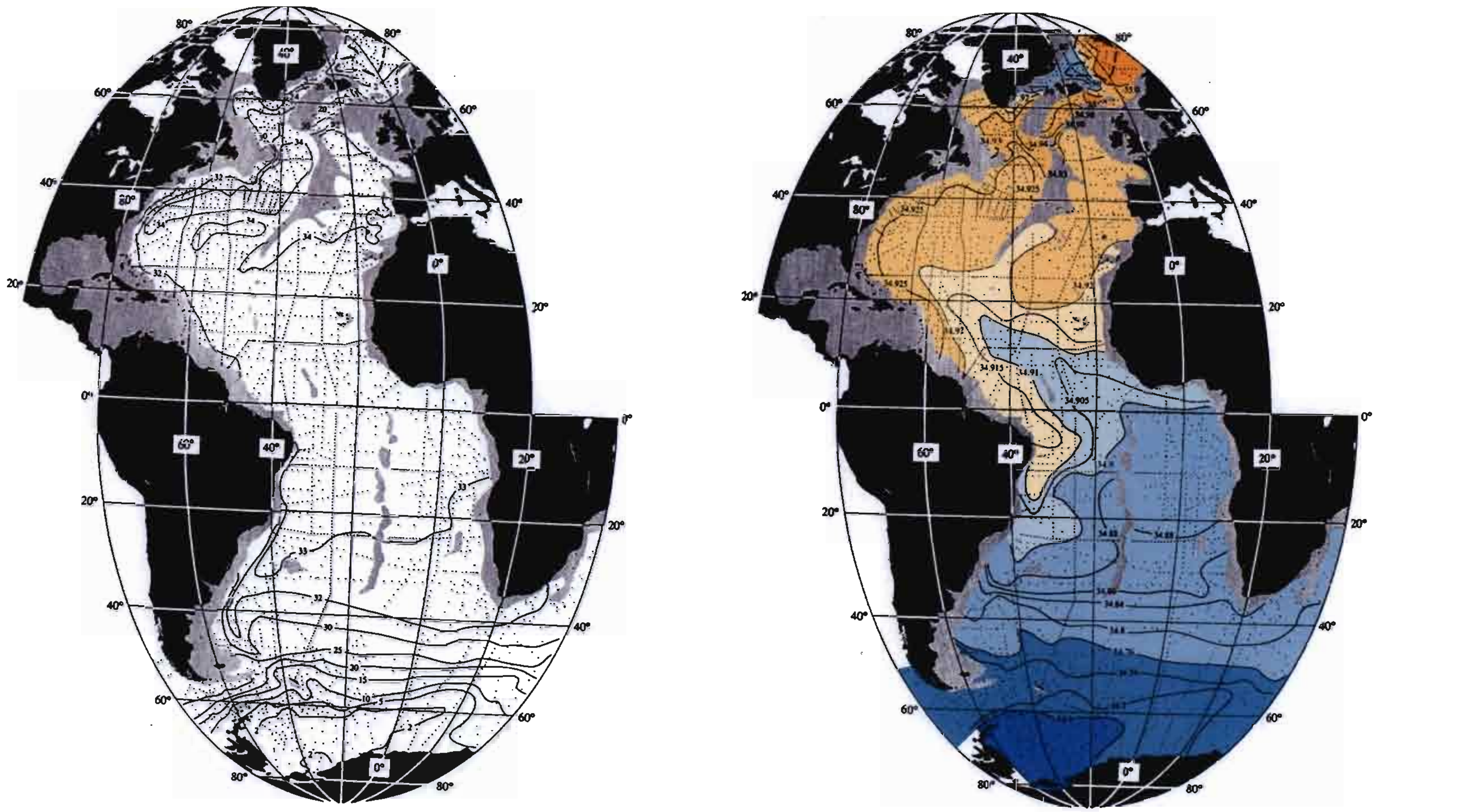

Fig. 31 Atlantic Ocean depth (hm) and salinity where σ_2 is 41.50 kg/m^3

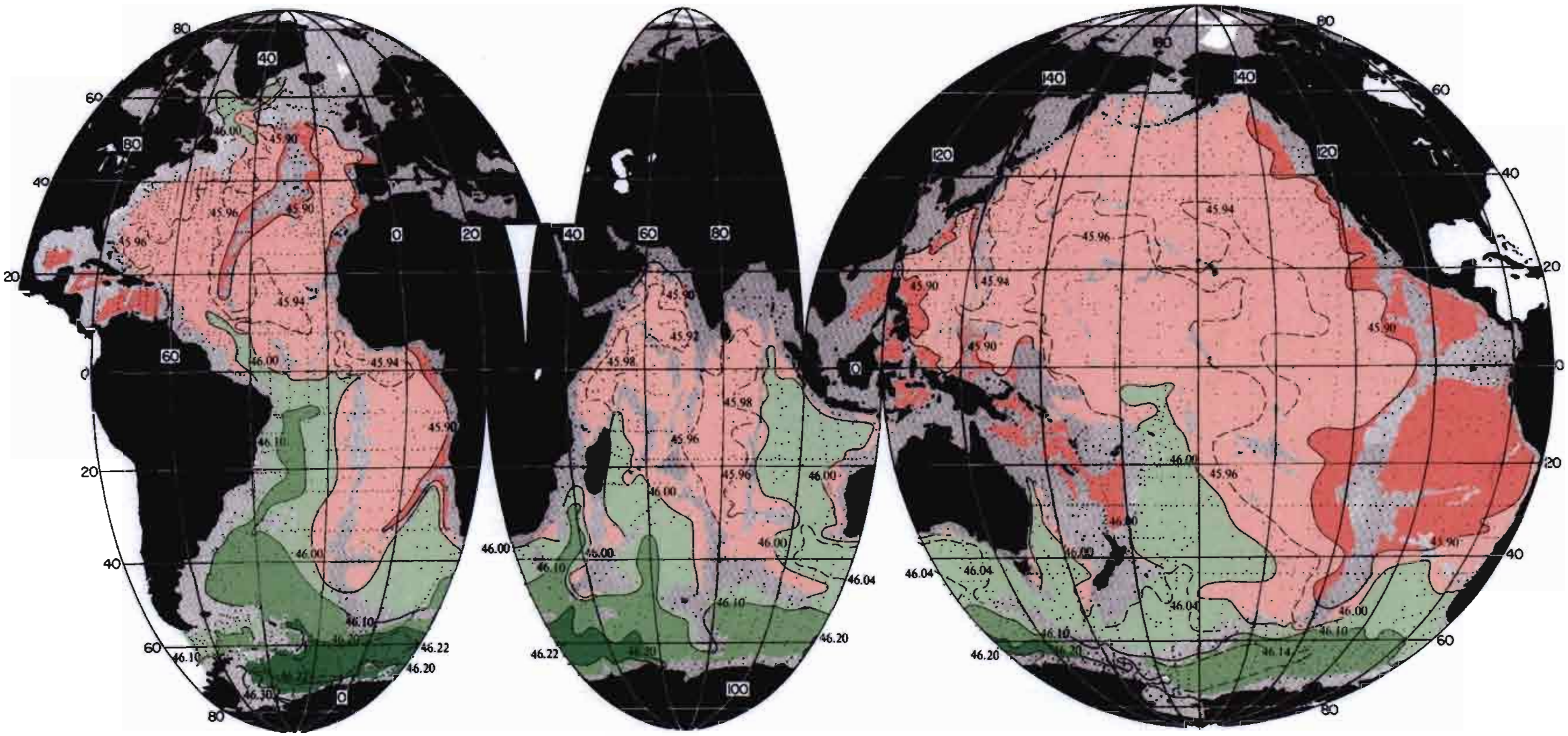

Fig. 32 World Ocean potential density at abyssal depths (greater than 3.5 km)

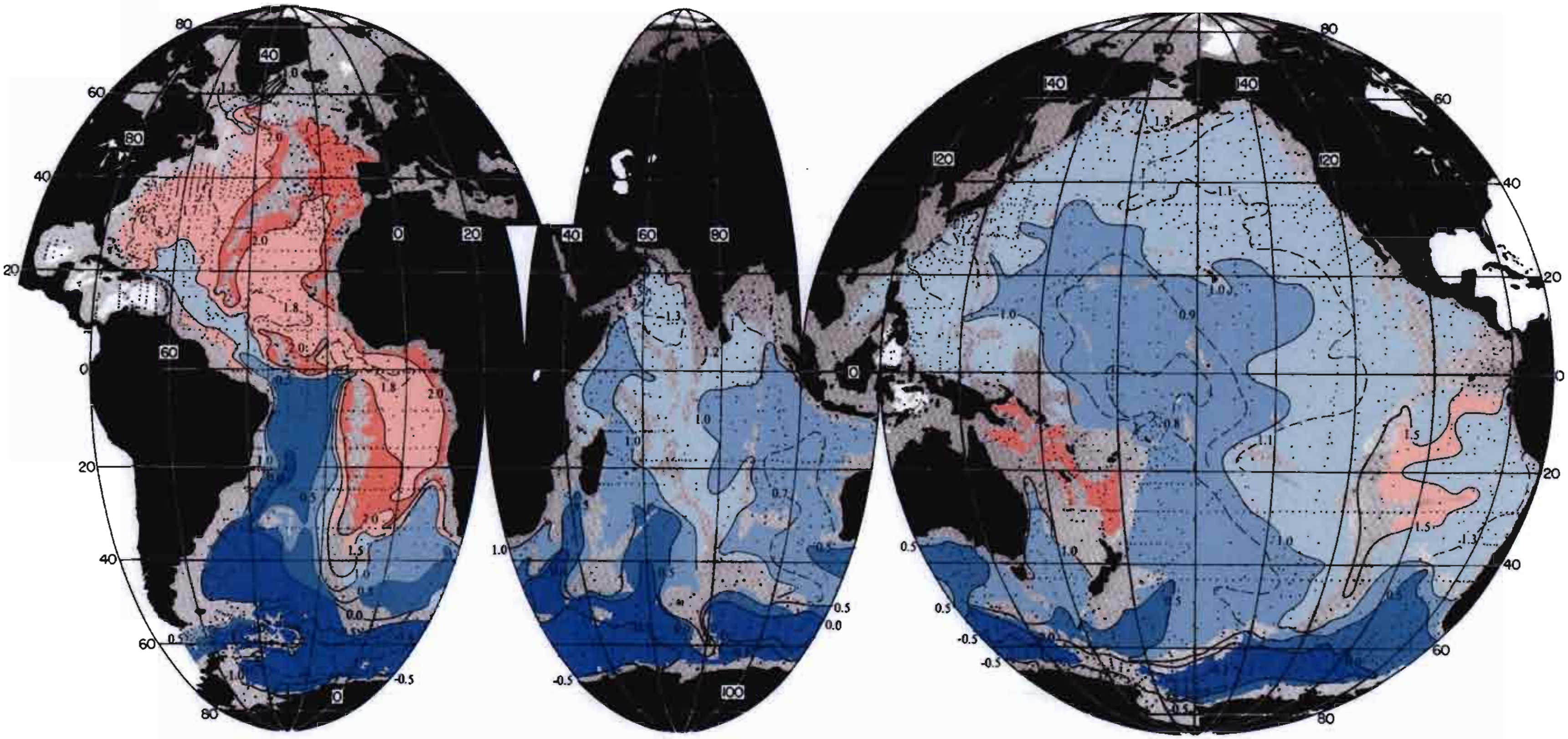

Fig. 33 World Ocean potential temperature at abyssal depths (greater than 3.5 km)

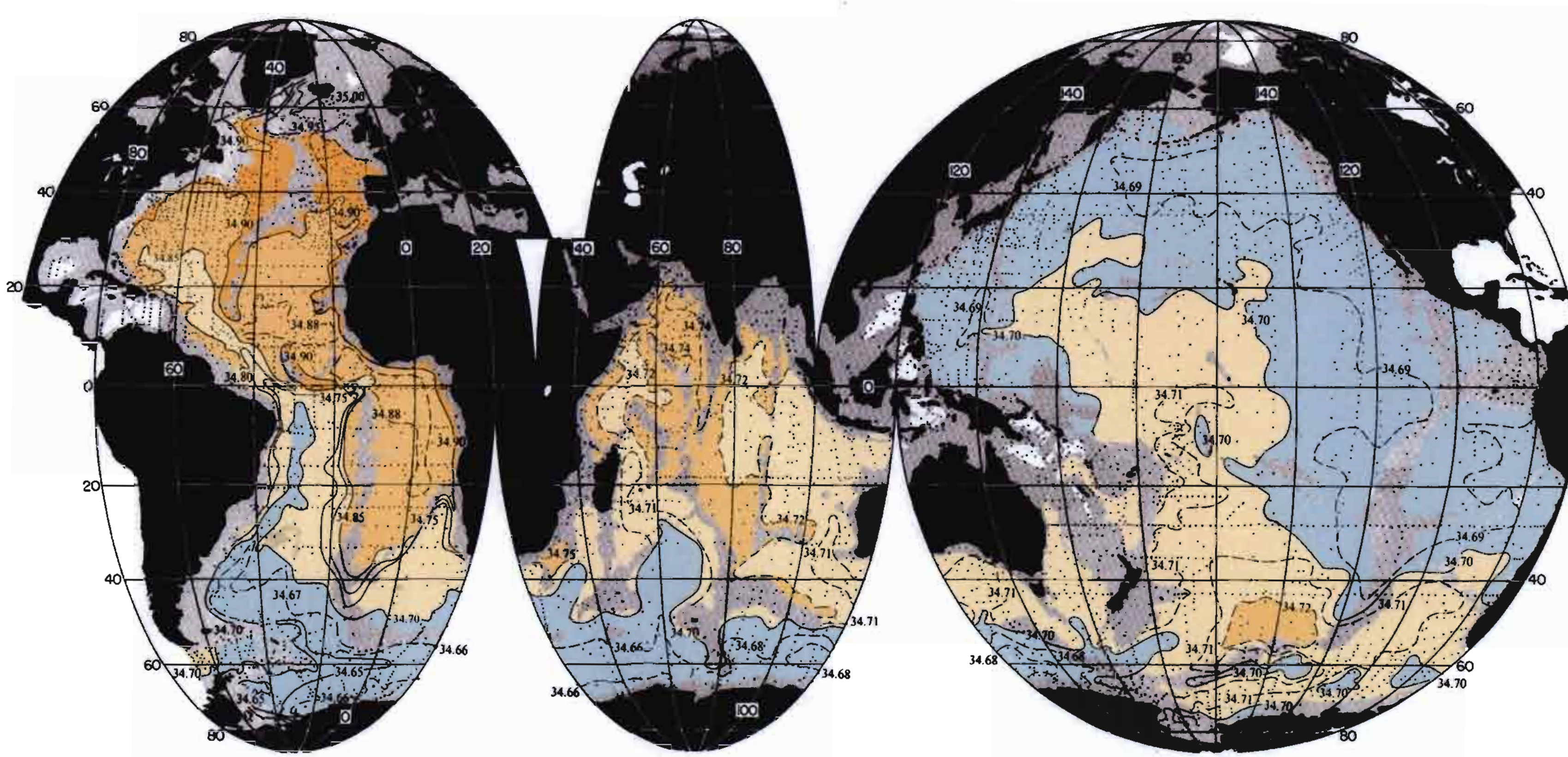

Fig. 34 World Ocean salinity at abyssal depths (greater than 3.5 km)