

RAPID RESPONSE PAPER

The geothermal heating of the abyssal subarctic Pacific Ocean

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Abstract—Recent deep CTD–O₂ measurements in the abyssal North Pacific along 175°W, 152°W, and 47°N indicate large-scale changes in the θ –S characteristics in the deepest kilometer of the water column. Geothermal heat flux from the abyssal sediments can be invoked as the agent for causing large-scale modification of abyssal temperatures (but not salinities) in the subarctic Pacific Ocean. East–west and north–south thermal age differences of about 100 years are inferred using a spatially uniform geothermal heat flux of $5 \times 10^{-2} \text{ W m}^{-2}$.

INTRODUCTION

THE abyssal waters of the northern North Pacific Ocean are among the oldest to be found anywhere, far from their origin in the Southern Ocean. They are thought to enter from a western boundary current in the Samoan Passage near 10°S, 169°W (REID and LONSDALE, 1974). MANTYLA and REID (1983) have prepared maps of the water mass characteristics of the abyssal waters which show a broad northward spreading of cold, oxygenated water in the central Pacific Basin to the west of the Christmas Ridge. From there, however, flow pathways to the subarctic Pacific are less definite. WARREN (1981a) has pointed out that poleward flow of the deep water into the central North Pacific is broad with no direct evidence for a deep western boundary current, although hints of one can be seen in the abyssal property maps of MANTYLA and REID (1983). The temperature and salinity characteristics of the abyssal waters are gradually altered by vertical mixing with the warmer, less saline overlying water. Bottom property distributions are consistent with a gradual erosion of the cold, salty abyssal properties towards the north and east.

Worthington's (1981) volumetric temperature and salinity census indicates that the abyssal North Pacific is quite uniform with the largest single bivariate θ –S class in the world ocean (1.1–1.2°C, 34.68–34.69‰). In this report recent data from several cruises will be presented which will show that measurable temperature–salinity variations exist in the northern North Pacific and that these changes are such that potential temperature increases toward the north and east with little or no change in salinity. Given that all abyssal water enters through the Samoan Passage, deep shifts in the θ –S curve must reflect either temporal variations in bottom water properties or some process acting to alter deep temperature without affecting the salinity. These changes are consistent with the notion that geothermal heat flux through the abyssal sediments alters the temperature (but not salinity) of the deepest kilometer of the subarctic Pacific Ocean.

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Other investigators (GORDON and GERARD, 1970; MANTYLA, 1975) have also sought to explain lateral variations in the near-bottom temperature field in terms of the geothermal heat flux. The inferences from their data were not entirely convincing though, because, as KNAUSS (1962) had pointed out, modification of the bottom water by vertical mixing with warmer, less saline water above is likely to be several times more significant than that due to heat flux from below, and will generally obscure that effect. The specific temperature variations that we wish to report, however, being unaccompanied by salinity variations, are not consistent with mixing with overlying water, and seem to be unambiguous evidence for a measurable effect of geothermal heating.

OBSERVATIONS

The spirit of this report is to illustrate evidence for geothermal heating of the abyssal subarctic Pacific. It is not meant to be an exhaustive data survey of the region which incorporates all of the recent data. Thus, widely spaced stations (Fig. 1) from four cruises have been selected and listed in Table 1. All data were collected with NBIS CTDs from the R.V. *Thomas Thompson* and R.V. *Thomas Washington*. Data from cruises TT-169, 117 were collected and processed by the WHOI CTD group, while TT-190 (TPS-47), and Marathon II (MII, aboard the *Washington*) were recently carried out by the PACODF group at Scripps. The stations broadly define the water mass characteristics in the central North Pacific from the subtropics to the Aleutians along the longitudes of 175 and 152°W, and from west to east at a latitude of 47°N. The final, calibrated temperature and salinity data are thought to be accurate to $\pm 0.002^\circ\text{C}$ and $\pm 0.002\text{‰}$ (with respect to the standard seawater used), respectively. All stations are bottom casts in water of depth about 5000 m or greater.

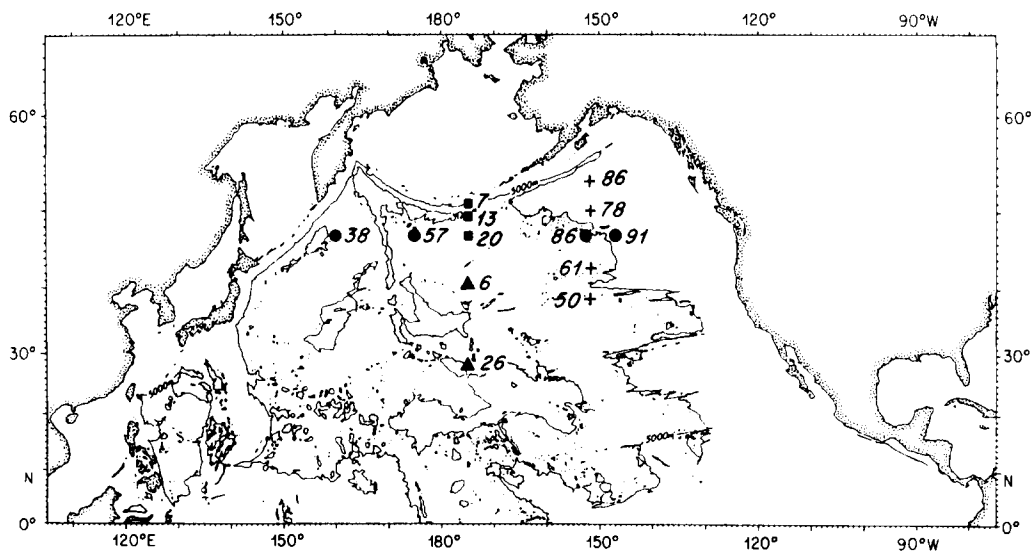


Fig. 1. Recent CTD stations along 47°N, 152°W, and 175°W in the North Pacific. Station locations and cruises are given in Table 1. Squares, triangles, circles and plusses denote cruises TT-169, TT-177, TPS-47 and MII, respectively.

Table 1. Station locations are plotted in Fig. 1 and θ - S diagrams are shown in Figs 2-4. They have been selected as a representative subset of a larger number of stations occupied on four different cruises

Cruise	Station No.	Date y/mo/day	Latitude	Longitude
TT-169	7	82.08.01	50°50.00'N	174°50.00'W
	13	82.08.03	49°15.25'N	174°50.76'W
	20	82.08.05	47°00.00'N	174°50.00'W
TT-177	6	83.10.31	40°30.00'N	174°59.99'W
	26	83.11.08	28°01.38'N	175°00.11'W
TPS-47	38	85.08.15	47°00.00'N	160°00.00'E
	57	85.08.21	47°00.00'N	174°56.80'E
	86	85.08.30	47°00.00'N </td <td>152°32.00'W</td>	152°32.00'W
	91	85.09.01	47°00.00'N	146°56.00'W
Marathon II	50	84.05.19	38°30.3'N	152°00.3'W
	61	84.05.24	42°55.4'N	151°58.3'W
	78	84.05.28	50°15.1'N	151°58.4'W
	86	84.05.31	53°41.1'N	151°58.9'W

Potential temperature-salinity diagrams for the selected stations have been plotted (Figs 2-4) which clearly show a spread in abyssal characteristics in the potential temperature range from 1.0 to 1.2°C. The water above 1.2°C has no detectable east-west variation on 47°N with some small north-south freshening on 175°W towards the Aleutians. The trend is for the abyssal water to warm as one moves from west to east and from south to north. North of Sta. 20, TT-169 (47°N, 174°50'W) and Sta. 84, MII (53°N, 152°W) the tendency is reversed and the θ - S curves rotate back towards more southerly

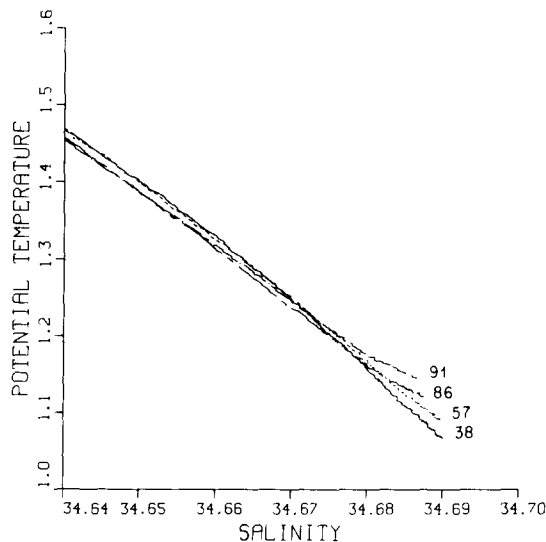


Fig. 2. Potential temperature vs salinity along 47°N. Stations are indicated on the figure and locations given in Table 1. These data are preliminary and subject to change once final calibrations are applied. These changes are likely to be at the 0.002‰ level and amount to horizontal shifts of the entire family of θ - S curves in the figure.

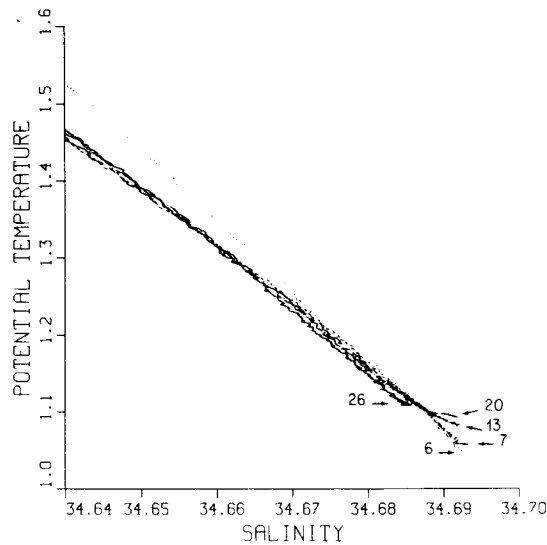


Fig. 3. As in Fig. 2, but along 175°W. While these data are "calibrated", it can be seen that salinity variations at the 0.001‰ level remain which are beyond our ability to calibrate to discrete water samples.

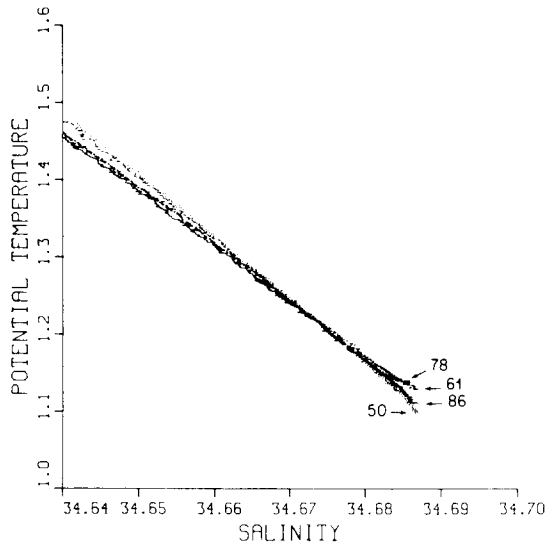


Fig. 4. As in Fig. 2, but along 152°W.

water properties below 1.1°C, 34.677‰ (Fig. 3) and 1.15°C, 34.68‰ (Fig. 4). Station 7, TT-169, which is in the Aleutian Trench, is similar to the deep θ -S structure of subtropical Sta. 6, TT-177. Station 86, MII, also reverts to the structure of Sta. 50, MII. In fact, Stas 7 and 86 lie in the deep eastward jet first identified by REED (1969), and water properties there are expected to be much more like those far to the south and west than immediately to the south (WARREN and OWENS, 1985).

The meridional structure of the deep θ - S relation can be illustrated by examining potential temperature changes for a fixed salinity. This has been done for the 152°W section (Fig. 5) for two salinity values: 34.685 and 34.687. These salinities are well within the region of the θ - S curves affected by geothermal heating. A positive anomaly indicates water that is warmer relative to the section as a whole. Consistent with the earlier remarks about the changes in the θ - S structure, a warm patch between 41 and 52°N is flanked by a zero-temperature-anomaly "plateau" to the south of 41°N and a narrow region of decreasing temperature anomaly towards the Alaskan Shelf. The large negative anomaly south of 30°N is to the north of the Hawaiian Range and shows the presence of relatively colder deep water which enters from the southeast around the Hawaiian Islands (e.g. WOOSTER and VOLKMANN, 1960; MANTYLA, 1975).

Abyssal water with the same salinity has a different potential temperature: the maximum difference is approximately 0.05°C on both 47°N and 152°W . The change of the deep θ - S relation is detectable at the highest temperatures ($\theta \sim 1.2^\circ\text{C}$) only in the eastern North Pacific. According to MANTYLA and REID (1983), abyssal waters with bottom potential temperatures in the range 1.0 - 1.2°C cover much of the subarctic Pacific. It is clear from the figures that WORTHINGTON'S (1981) largest bivariate θ - S class of 1.1 - 1.2°C , 34.68 - 34.69‰ has significant structure to it. A reasonable explanation for the observed variations is that the relatively slow abyssal circulation acquires heat (but not salt) from the sediments. Temperature then becomes a tracer which age-dates the water. The "oldest" water is therefore the warmest.

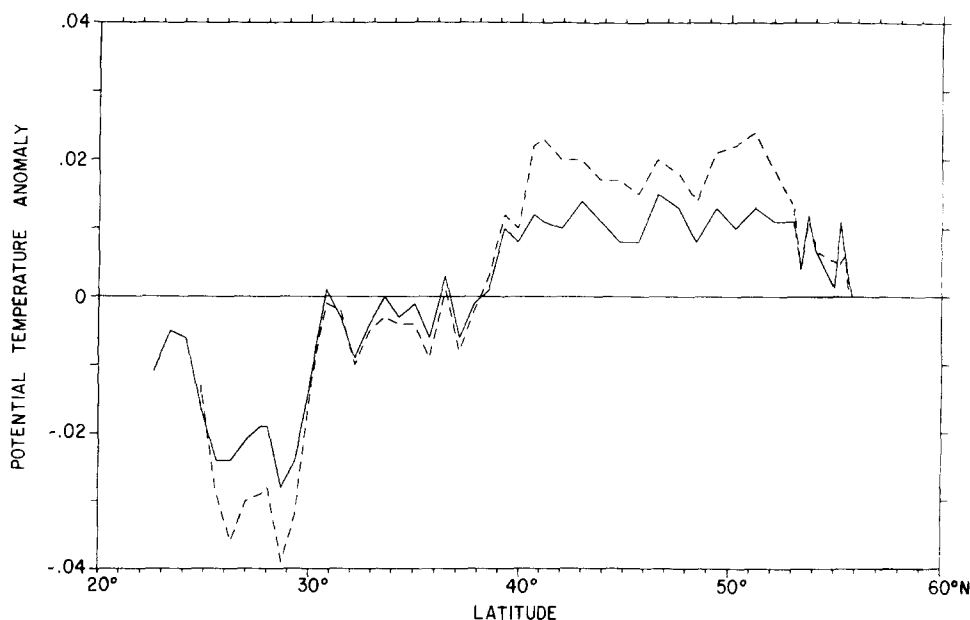


Fig. 5. Potential temperature anomaly at salinities 34.685 (solid line) and 34.687‰ (dashed line) along 152°W . An average θ - S relation was computed for 0.01°C intervals using all stations along 152°W from which it was determined that the average salinity at 1.2°C was 34.676‰ , at a point on the θ - S curve above the region affected by the deep geothermal heating and where most of the θ - S curves appeared to cross. Each station's salinities were therefore adjusted uniformly to be 34.676‰ at 1.2°C . The average θ - S relation was then recomputed for 0.001‰ intervals and the anomaly at a given salinity was recomputed.

THEORETICAL CONSIDERATIONS

Before one can utilize temperature as a tracer, it is necessary to demonstrate that geothermal heating does not affect the dynamics of the abyssal circulation. If the dynamics are steady, hydrostatic, geostrophic, and β -plane, then

$$\begin{aligned} \rho f u &= -p_y \\ -\rho f v &= -p_x \\ 0 &= -g\rho - p_z \\ (u\rho)_x + (v\rho)_y + (w\rho)_z &= -F_z^b \\ \rho &= \bar{\rho}(1 - \alpha_0 \theta + \alpha_s S), \end{aligned} \quad (1)$$

where the position vector (x, y, z) is oriented east, north, up and the corresponding velocity vector is (u, v, w) , f is the Coriolis parameter, α_0 and α_s , the thermal expansion and haline contraction coefficients, respectively, and F^b represents the vertical flux of buoyancy by small-scale mixing. Horizontal mixing has been neglected and a (constant) reference density is given by $\bar{\rho}$. In abyssal circulation models, horizontal gradients of temperature and salinity are often neglected. This yields a steady balance (WYRTKI, 1961; MUNK, 1966) between vertical advection and diffusion. While still holding in the deep ocean, this simplification is not valid in the bottom kilometer where horizontal temperature and density gradients (due to a heat flux from the sediments) are found.

The momentum and density equations can be combined to give a vorticity equation.

$$\rho\beta v = f(F^b + \rho w)_z, \quad (2)$$

where $\beta \equiv f_y$.

If this vorticity equation is integrated from the ocean bottom ($w = 0$) to a depth z_1 where vertical advection and diffusion balance, then for small density differences, the vorticity equation can be written

$$\beta \int_{z_b}^{z_1} v dz = f [w(z_1) - F^b(z_b)/\bar{\rho}]. \quad (3)$$

This is a slightly modified version of the STOMMEL and ARONS (1960) model of the abyssal circulation driven by localized sinking and deep upwelling. The additional term represents the modification due to a buoyancy (heat) flux into the ocean from the sediments. The buoyancy driving arises because, at this stage, the Boussinesq approximation (SIEGEL and VERONIS, 1960) has not been invoked. This term is, however, very small and can be estimated by expressing the buoyancy flux in terms of its constituents:

$$F^b = -\frac{\alpha_0}{C_p} F^Q + \alpha_s F^S. \quad (4)$$

The ratio of the buoyancy driving to the upwelling term is

$$\frac{F^b(z_b)}{\bar{\rho}w(z_1)} = -\frac{\alpha_0 F^Q(z_b)}{\bar{\rho}C_p w(z_1)} \sim 0 \quad (10^{-5}),$$

after inserting "typical" values for the above of

$$\begin{aligned} \alpha_0 &\sim 2 \times 10^{-4} (\text{°C})^{-1} \\ \bar{\rho} C_p &\sim 4 \times 10^6 \text{ J m}^{-3} (\text{°C}) \\ F^Q(z_b) &\sim 5 \times 10^{-2} \text{ W m}^{-2} \\ w &\sim 2 \times 10^{-7} \text{ m s}^{-1}. \end{aligned}$$

Thus to lowest order, a geothermal heat flux into the ocean of 50 mW m^{-2} , which is representative of the North Pacific (LANGSETH and VON HERZEN, 1970), does not significantly affect the vorticity balance of the abyssal circulation. Furthermore, the Boussinesq approximation can be made with no loss of important dynamics. Hence (1) becomes

$$\begin{aligned} \bar{\rho} f u &= -p_y \\ -\bar{\rho} f v &= -p_x \\ 0 &= -g\rho - p_z \\ u\rho_x + v\rho_y + w\rho_z &= -F_z^h \\ \rho &= \bar{\rho} (1 - \alpha_0\theta + \alpha_s S) \end{aligned} \quad (1')$$

and the salinity-potential temperature equations can be written as

$$\begin{aligned} \bar{\rho} (uS_x + vS_y + wS_z) &= -F_z^T \\ \bar{\rho} C_p (u\theta_x + v\theta_y + w\theta_z) &= -F_z^Q \end{aligned} \quad (5)$$

Bottom heating can produce a vertical shear and a turning of the horizontal velocity vector. If the horizontal velocities are written in polar form as $(u, v) = s (\cos \phi, \sin \phi)$, then following SCHOTT and STOMMEL (1978), the vertical change of direction can be expressed using (1') as

$$fs^2\phi_z = \frac{g}{\bar{\rho}} (w\rho_z + F_z^h) \quad (6)$$

which is positive owing to bottom heating. The turning will vanish in the event of a 1-dimensional Munk-Wyrtki balance and is inversely proportional to the square of the current speed. If in the lower kilometer, the 1-dimensional balance is upset, then an upper bound on the amount of turning can be estimated by ignoring the $w\rho_z$ term and expressing the right-hand side as

$$\frac{g}{\bar{\rho}} (w\rho_z + F_z^h) \leq \frac{g\alpha_0 F^Q}{\bar{\rho} C_p h}.$$

If the vertical velocity is $2 \times 10^{-7} \text{ m s}^{-1}$ (CRAIG, 1969) at a height of 3 km above the bottom, then the vertically averaged speed, s , at greater depths can be estimated (STOMMEL and ARONS, 1960) to be 0.7 mm s^{-1} at latitude 45°N in mid-Pacific. The maximum direction change over a vertical distance h would then be

$$\Delta\phi = \frac{g\alpha_0 F^Q}{f\bar{\rho} C_p s^2} \approx 30^\circ$$

which is not likely to be noticeable in property distributions. However, because the amount of turning varies inversely with the current speed, the effect might be significantly larger in lower latitudes, where the speed (as well as f) should be smaller. On the other hand, regional variation in bottom slope can increase the flow speed severalfold (e.g. WARREN and OWENS, 1985) over the flat-bottom estimate above, so the real turning may, in fact, be much less than suggested. It is also unclear how the vertically averaged speed relates to the near-bottom speed, which is actually the pertinent quantity for this estimate, but is more difficult to infer.

It therefore appears that geothermal heating will have a negligible effect on the vorticity balance and, with less certainty, the direction of the abyssal circulation. To some degree of approximation, temperature can be considered a passive tracer of the flow near the bottom.

To calculate the temperature change due to geothermal heat flux, resolve the temperature field into an interior field θ_i , which is determined as if the bottom flux were zero, and a near-bottom anomaly, θ_a , which absorbs the geothermal flux and decays (with its associated diffusive flux F_a^Q) to zero some distance z_1 above the bottom (at depth z_b). Then θ_a (as well as θ_i and their sum $\theta = \theta_i + \theta_a$) satisfies a version of (5):

$$\bar{\rho} C_p (u\theta_{ax} + v\theta_{ay} + w\theta_{az}) = -F_a^Q. \quad (7)$$

Assume no horizontal variation in z_b and z_1 , and no variation in horizontal current between z_b and z_1 , so that $w_z = \text{constant}$ and $w(z_b) = 0$. With the definitions

$$\bar{\theta}_a \equiv \frac{1}{h} \int_{z_b}^{z_1} \theta_a \, dz, \quad h \equiv z_1 - z_b,$$

the vertically integrated form of (7) is

$$h(u\bar{\theta}_{ax} + v\bar{\theta}_{ay}) - w(z_1)\bar{\theta}_a = \frac{F_a^Q(z_b)}{\bar{\rho} C_p}. \quad (8)$$

where $\theta_a(z_1)$ and $F_a^Q(z_1)$ have been set to zero according to the definition of θ_a , and $F_a^Q(z_b)$ is now the geothermal heat flux.

Following a fluid column moving with velocity (u, v) , temperature changes can be expressed as

$$\frac{d\bar{\theta}_a}{dt} - \frac{w(z_1)}{h} \bar{\theta}_a = \frac{F_a^Q(z_b)}{\bar{\rho} C_p h}. \quad (9)$$

The ratio of the second to the first term in (8) or (9) can be shown to be $0(\beta L/f) < 1$. Hence, the solution of (9), for an initial condition of $\bar{\theta}_a = 0$ and consistent with the β -plane approximation, can be written

$$\bar{\theta}_a = \frac{F_a^Q(z_b)t}{\bar{\rho} C_p h} \left[1 + \frac{1}{2} \frac{w}{h} t + \dots \right]. \quad (10)$$

Again for $F_a^Q(z_b)$ of 50 mW m^{-2} and a layer depth of 1000 m, and neglecting the terms in w , we estimate that for a time of 100 years, the average temperature increase in abyssal water is 0.04°C .

In performing the time-integration, it has been assumed that following a fluid column of thickness h , the geothermal heat flux and the upwelling velocity are constant. The heat flux is not, in fact, constant over the ocean floor, but in the central basin of the North Pacific, and specifically in the region of our stations, its spatial variations are neither large nor systematic except over the Aleutian Island Arc and Trench, where values may be twice as high as the world average (LANGSETH and VON HERZEN, 1970). For the basins of the Pacific generally (away from ridges and trenches), LANGSETH and VON HERZEN (1970) report an average value of 54 mW m^{-2} with a standard deviation of 16 mW m^{-2} , which suggests an upper limit uncertainty of 30% in our estimate of temperature change after 100 years (or, conversely, of time taken to achieve a change of 0.04°C) due to variations in heat flux. Of course, the actual uncertainty must be a good deal smaller because the effects of random geographical variations in heat flux are integrated out along the trajectory of a moving fluid column.

At mid-depths, a plausible range in estimates of the upwelling velocity is $1\text{--}3 \times 10^{-7} \text{ m s}^{-1}$ (see below). Within 1 km of the bottom, however, the upwelling must be much reduced, perhaps to only one-third as large, so for a transit time of 100 years the correction terms for w in (10) can be safely assumed to be $<20\%$ and therefore unimportant, whatever its likely spatial variation.

DISCUSSION

Data have been presented showing up to 0.05°C temperature differences in waters with the same salinity. If potential temperature is taken to be a passive tracer, then estimated geothermal heat flux following fluid columns will heat up the abyssal waters at a rate of $0.04^\circ\text{C} (100 \text{ y})^{-1}$ when the heat input is distributed over a layer 1000 m thick. Examination of individual casts indicates that the deep water-mass anomalies are indeed about 1 km in thickness. This can be illustrated in Fig. 6 where two stations (TT-169, 7, 13) have been selected and potential temperature, salinity and density (σ_5) plotted for depths below 4000 m. The two stations have been "aligned" by shifting Sta. 13 upward by ~ 200 m so that temperatures and salinities are the same at 4000 m. Effects of heating have made the deep potential temperature in the lower kilometer at Sta. 13 higher than at Sta. 7, which is over the Aleutian Trench. Note that the salinity profiles are nearly the same and thus most of the density difference in the "aligned" profiles, and up to 50% of that before alignment, is a result of variations in the deep potential temperature.

Because salinity increases with depth in the deep North Pacific, heating from the bottom need not destabilize the bottom water but merely leads to upward diffusion of a positive temperature anomaly without disturbing the salinity field. This stratification contrasts with that in the deep Atlantic and Indian oceans, where salinity decreases with depth, and where geothermal heating might cause convective overturning and stirring of the bottom water and a much less easily discernible signature of the heat flux.

Penetration of the geothermal warming to about 1000 m above the bottom in a time interval of 100 years seems consistent with purely diffusive heat transfer. If a fluid column moving uniformly with speed s is subject to a heat flux F^Q at the bottom, then the resulting temperature anomaly θ_a would be distributed by vertical diffusion alone as

$$\theta_a = \frac{2F^Q}{\rho C_p} \sqrt{\frac{l}{ks}} i^1 \text{erfc} \frac{z}{2} \sqrt{\frac{s}{kl}}, \quad (11)$$

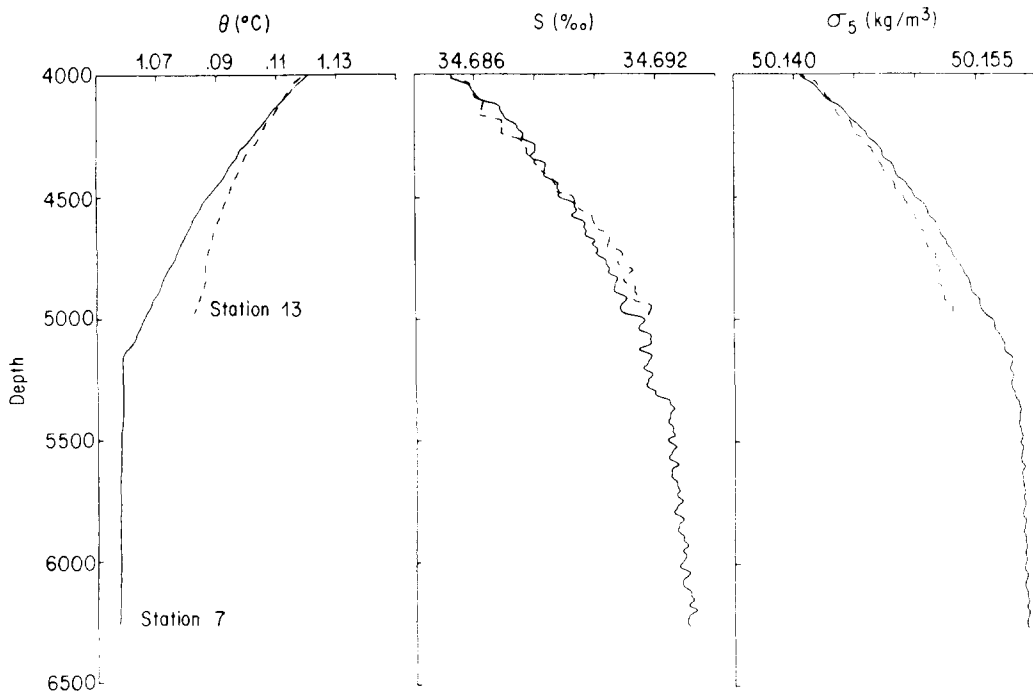


Fig. 6. Profiles of potential temperature (left), salinity (center), and potential density (σ_5 , right) for TT-169 Stas 7 (solid line) and 13 (dashed line). The latter has been shifted vertically ~ 200 m to illustrate the vertical, rather than horizontal differences in properties. Station 7 is over the Aleutian Trench and has a nearly adiabatic temperature below the sill depth of the trench.

where l is the distance along a streamline from the inception of heating, z is the height above the bottom, k is the (constant) thermal diffusivity, and $i^1\text{erfc}$ is the first integral of the complementary error function. If the "penetration" height, z_p , is defined as the value of z for which θ_a is 9% of its bottom value, then $z_p = 2 \sqrt{s/k}l$. For z_p to be 1000 m after $l/s = 100$ years, then $k = 0.8 \text{ cm}^2 \text{ s}^{-1}$, which is a typical estimate of deep ocean diffusivity. (If the time l/s is uncertain by 30% owing to spatial variations in heat flux, then the same uncertainty applies to this estimate of k .)

It must be remembered that relations (9)–(11) were derived following fluid motion. We have no reason to assume that flow lines will pass between any two of our observations. Therefore (10) should be used as an index of relative age only. The "oldest" water is in the northeastern Pacific, but not immediately against the northern boundary. The maximum "age" difference is roughly 100 years. As may be inferred from Figs 2 to 4, the water in the central North Pacific has a similar relative age to that in the northwest (Sta. 38) and immediately south of the Aleutians [Stas 7 and 86 (MII)]. South of the Aleutian trench and extending eastward to North America is a tongue of water 50–100 years "older" than to the south or west.

A circulation model driven by deep upwelling and including idealized bottom topography (WARREN and OWENS, 1985) seems to rationalize REED'S (1969) eastward jet, in company with a westward boundary current along the Aleutians just to the north of the eastward flow. The deep water of both currents in the model is "young" relative to that

just to the south, and therefore the circulation scheme will produce a local maximum in "age" south of these strong zonal currents. For an upwelling velocity at the "top" of the deep water of $2 \times 10^{-7} \text{ m s}^{-1}$, this predicted "age" difference is about 150 years, which compares reasonably well with our data. This model "age" comparison, as well as some previous arguments, depends on the assumed value of the deep upwelling velocity, which is not measurable and has only been estimated from fits of data to various models, of uncertain relevance. Some note of the range of plausible estimates may therefore be pertinent. At one extreme, CRAIG (1969) analyzed Pacific radiocarbon and stable carbon data in terms of a vertical advection-diffusion model, and calculated the required upwelling velocity at mid-depths to be $2.2 \times 10^{-7} \text{ m s}^{-1}$. At the other extreme, WARREN (1981b) inserted transport estimates for deep western-boundary currents at latitude 43°S into a Stommel-Arons circulation model, and inferred an average upward velocity near 2000 m over the Pacific of $0.8 \times 10^{-7} \text{ m s}^{-1}$. It seems likely, therefore, that the mid-depth upwelling velocity will be found in the range $1\text{--}3 \times 10^{-7} \text{ m s}^{-1}$, though there is no reason to suppose that it should be uniform geographically. The corresponding range of "age" differences given by the Warren-Owens model is then about 100–300 years.

If the upwelling velocity is only $1 \times 10^{-7} \text{ m s}^{-1}$, our previous flat-bottom estimate of the vertical turning of the horizontal velocity vector due to geothermal heat flux would be four times larger, and not consistent with our present model of near-bottom water moving as vertical columns. As noted before, however, it seems probable that topographic variation will force mean flow speeds large enough to preclude much turning. Certainly the mean deep speeds reported by WARREN and OWENS (1985) at a longitude of 175°W are too large to allow any noticeable effect.

If potential temperature can date bottom water, how do these ages compare with those estimated from oxygen depletion and radiocarbon decay? Oxygen samples were collected for all of the stations identified in Fig. 1. A plot of oxygen concentration against salinity (not potential temperature!) shows (Fig. 7) that the deep water at Sta. 91, TPS-47, is approximately 0.25 ml l^{-1} lower in dissolved oxygen concentration than that at Sta. 38 for comparable salinities between 34.64 and 34.70‰. Thus there is an oxygen decrease of 0.25 ml l^{-1} from west to east across the deep subarctic Pacific. In combination with a temperature "age" difference of 100 years, this would be consistent with an *in situ* oxygen-consumption rate of $0.0025 \text{ ml l}^{-1} \text{ y}^{-1}$: close to the "best-fit" estimate of $0.002 \text{ ml l}^{-1} \text{ y}^{-1}$ by ARONS and STOMMEL (1967) and KUO and VERONIS (1973) in their tests of deep-circulation models.

Large-volume sampling for radiocarbon was not carried out on any of the cruises discussed here, but STUIVER *et al.* (1983) have used the GEOSECS data to examine meridional variations in radiocarbon activity ($\Delta^{14}\text{C}$) in the world ocean. They concluded from station-averaged values below 1500 m and from minimum-activity values near 2500 m, that the oldest water in the deep ocean was in the North Pacific near latitude 40°N , distinctly *south* of the Aleutian Trench, rather than against the northern boundary. This finding would confirm our own inference, except that the northward-increasing activities north of latitudes $35\text{--}40^{\circ}\text{N}$ correlate with decreasing concentrations of total carbon dioxide (ÖSTLUND and STUIVER, 1980), and there seems to be *no* significant meridional trend in the absolute radiocarbon concentrations ($\Sigma^{14}\text{C}$) calculated from these activities and carbon-dioxide concentrations. The stations are far too few to prove lateral uniformity of absolute radiocarbon concentrations in the deep subarctic Pacific, but the available measurements do not reveal an age decrease toward the Aleutians.

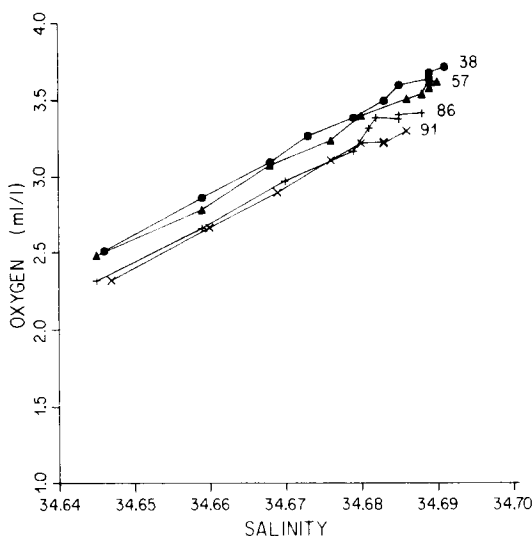


Fig. 7. Four stations from TPS-47 (Fig. 2) have been selected and the variation of oxygen plotted vs salinity. In this presentation, the bottom water has the highest oxygen and salinity and is at the upper right of the plot. A consistent trend of lower oxygen for a particular salinity can be traced across the section from west to east at nearly all bottle depths.

On the other hand, the GEOSECS stations in latitudes 30–35°N constitute a zonal section of radiocarbon in the *subtropical* North Pacific. The most prominent feature of the vertically averaged values below 1500 m is an eastward decrease in absolute radiocarbon concentration [as calculated from the data listed by ÖSTLUND and STUIVER (1980)], such that the vertically averaged concentrations east of the Emperor Seamounts are about 1% lower than those to the west. If this decrease were due to radiocarbon decay alone, then for a decay rate of 1% every 83 years (STUIVER *et al.*, 1983), its magnitude would be consistent with our estimate of east–west age difference farther north. Of course, the distribution of absolute concentration is affected by much more than its decay rate (CRAIG, 1969), and in any case we do not know whether this zonal trend exists in the subarctic Pacific.

The oxygen and radiocarbon evidence may perhaps be conformable to our “age” inference, but while each type of evidence offers advantages not presented by the temperature variations, neither seems so decisive as the latter. Oxygen data provide useful evidence throughout the water column, but the forcing term for the oxygen distribution, local consumption rate, is not known from direct measurement and has only been estimated by fitting data to simple models including free parameters. Radiocarbon data also may be valid age indicators over much of the water column, and the forcing term, decay rate, is very well known, but the interpretation of radiocarbon concentrations is model-dependent (CRAIG, 1969) and not straightforward; moreover, radiocarbon data are (and probably always will be) much sparser than temperature data. Effects of geothermal heating, on the other hand, are confined to the bottom water, but the forcing term has been well measured directly, and in the example we have discussed, the effects and their interpretation seem unambiguous. We suggest therefore that in some circumstances, temperature variations due to geothermal heat flux can be isolated and used as clear-cut, quantitative tracers of near-bottom circulation.

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REFERENCES

- ARONS A. B. and H. STOMMEL (1967) On the abyssal circulation of the world ocean—III. An advection-lateral mixing model of the distribution of a tracer property in an ocean basin. *Deep-Sea Research*, **14**, 441–457.
- CRAIG H. (1969) Abyssal carbon and radiocarbon in the Pacific. *Journal of Geophysical Research*, **74**, 5491–5506.
- GORDON A. L. and R. D. GERARD (1970) North Pacific bottom potential temperature. In: *Geological investigations of the North Pacific*, J. D. HAYS, editor, GSA Memoir 126, pp. 23–39.
- KNAUSS J. A. (1962) On some aspects of the deep circulation of the Pacific. *Journal of Geophysical Research*, **67**, 3943–3954.
- KUO H. H. and G. VERONIS (1973) The use of oxygen as a test for an abyssal circulation model. *Deep-Sea Research*, **20**, 871–885.
- LANGSETH M. G., Jr and R. P. VON HERZEN (1970) Heat flow through the floor of the world oceans. In: *The sea*, Vol. 4, Part I, A. E. MAXWELL, editor, Wiley-Interscience, New York, pp. 299–352.
- MANTYLA A. W. (1975) On the potential temperature in the abyssal Pacific Ocean. *Journal of Marine Research*, **33**, 341–354.
- MANTYLA A. W. and J. L. REID (1983) Abyssal characteristics of the world ocean waters. *Deep-Sea Research*, **30**, 805–833.
- MUNK W. H. (1966) Abyssal recipes. *Deep-Sea Research*, **13**, 707–730.
- ÖSTLUND H. G. and M. STUIVER (1980) GEOSECS Pacific radiocarbon. *Radiocarbon*, **22**, 25–53.
- REID R. K. (1969) Deep water properties and flow in the central North Pacific. *Journal of Marine Research*, **27**, 24–31.
- REID J. L. and P. F. LONSDALE (1974) On the flow of water through the Samoan Passage. *Journal of Physical Oceanography*, **4**, 58–73.
- SCHOTT F. and H. STOMMEL (1978) Beta spirals and absolute velocities in different oceans. *Deep-Sea Research*, **25**, 961–1010.
- SPIEGEL F. A. and G. VERONIS (1960) On the Boussinesq approximation for a compressible fluid. *Astrophysics Journal*, **131**, 444–447.
- STOMMEL H. and A. B. ARONS (1960) On the abyssal circulation of the world ocean—II. An idealized model of circulation pattern and amplitude in oceanic basins. *Deep-Sea Research*, **6**, 217–233.
- STUIVER M., P. D. QUAY and H. G. ÖSTLUND (1983) Abyssal water carbon-14 distribution and the age of the world oceans. *Science*, **219**, 849–851.
- WARREN B. A. (1981a) Deep circulation of the world ocean. In: *Evolution of physical oceanography*, B. A. WARREN and C. WUNSCH, editors, MIT Press, Cambridge, MA, pp. 6–41.
- WARREN B. A. (1981b) Transindian hydrographic section at latitude 18°S: Property distributions and circulation in the South Indian Ocean. *Deep-Sea Research*, **28**, 759–788.
- WARREN B. A. and W. B. OWENS (1985) Some preliminary results concerning deep northern-boundary currents in the North Pacific. *Progress in Oceanography*, **14**, 537–551.
- WOOSTER W. S. and G. H. VOLKMANN (1960) Indications of deep Pacific circulation from the distribution of properties at five kilometers. *Journal of Geophysical Research*, **65**, 1239–1249.
- WORTHINGTON L. V. (1981) The water masses of the world ocean: some results of a fine-scale census. In: *Evolution of physical oceanography*, B. A. WARREN and C. WUNSCH, editors, MIT Press, Cambridge, MA, pp. 42–69.
- WYRTKI K. (1961) The thermohaline circulation in relation to the general circulation in the oceans. *Deep-Sea Research*, **8**, 39–64.