The North Pacific: A Global-Scale Estuary

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ABSTRACT

The atmospheric net flow of water from the Atlantic to the Pacific Ocean is supposed to maintain the salinity difference between the two oceans. Assuming the existence of a subsurface level of no horizontal pressure gradient in the ocean, the mean sea level in the northern Pacific must be higher than in the Arctic Ocean. This mean sea level difference is supposed to drive the observed mean flow through the Bering Strait.

The estimated flow of freshwater through the Bering Strait is approximately equal to the estimated atmospheric net flow of water from the North Atlantic to the North Pacific. This justifies the formulation of a simple estuary model for the North Pacific in which the "brackish" water exits through the Bering Strait. The salinity difference between the two oceans is shown to be controlled by the topography of the Bering Strait. The estuary model gives residence times for water in the upper layer ($\sim$1000 m thick) of approximately 1000 years and in the lower layer of approximately 4000 years.

1. Introduction

On a global scale, the North Pacific acts like an estuary. There is a net atmospheric flow of water from the North Atlantic to the North Pacific, essentially across Central America, see e.g., Dietrich (1963) and Weyl (1968). This flow tends to 1) decrease the salinity of the North Pacific waters and 2) increase the salinity of the North Atlantic waters. These tendencies are opposed by advective transports in the ocean and something resembling a steady-state salinity distribution is achieved. After mixing with underlying denser water, the lighter Pacific surface water flows toward the Atlantic Ocean while denser Atlantic water flows at greater depths toward the Pacific. From such a simple picture one would expect that there is a horizontal pressure gradient toward the Atlantic in the surface layers, driving the surface flow. At greater depths, the horizontal pressure gradient should reverse, driving deep water from the Atlantic into the Pacific. In an intermediate depth interval, the horizontal pressure gradient should be weak. A definition sketch is given in Fig. 1. If we know the depth of the zero horizontal pressure gradient, we can calculate the hydrostatic sea level difference between the North Atlantic and the North Pacific. Because the envisaged source–sink flow takes place in oceans with complicated topographies on a rotating sphere, the induced current pattern is generally rather complex.

The North Pacific has two connections with the North Atlantic: a long one, via the South Pacific and the South Atlantic, which is deep and wide, and a very short one, via the shallow and narrow Bering Strait. In this paper it is argued that the freshwater cycle between the North Pacific and the North Atlantic is closed by the transport of water of low salinity through the Bering Strait.

In a typical strait like, e.g., the Gibraltar Strait or a fjord mouth, the net transport through the strait is equal to the volume flow from river runoff plus precipitation minus evaporation at the fjord surface. Superposed on the net transport there is often a two-layer circulation induced by mechanical mixing and/or thermohaline processes. This may be one or more orders of magnitude larger than the net transport.

Because of the existence of two connections to the "sea" (i.e., the North Atlantic) the "estuarine" circulation of the North Pacific seems to take an unusual form. The shallow Bering Strait only permits exchange of surface water. Therefore, the baroclinic (estuarine) circulation caused by the atmospheric net flow of freshwater from the North Atlantic seems to be separated. The less saline surface waters take the direct pass to the Atlantic through the Bering Strait while the "salt compensation" current must come in from the south (via the South Pacific); see Fig. 1.

The hydrostatic sea level difference between the North Pacific and the North Atlantic is calculated by assuming, somewhat arbitrarily, that the level of zero horizontal pressure gradient is situated at a depth of 1100 m. The sea level difference drives the mean flow in the Bering Strait.

The concept of the North Pacific as a global-scale estuary is developed at the end of this note. A steady-state estuary model for the North Pacific is proposed. It is demonstrated that the magnitude of the salinity difference between the two oceans is probably topographically controlled by the Bering Strait.

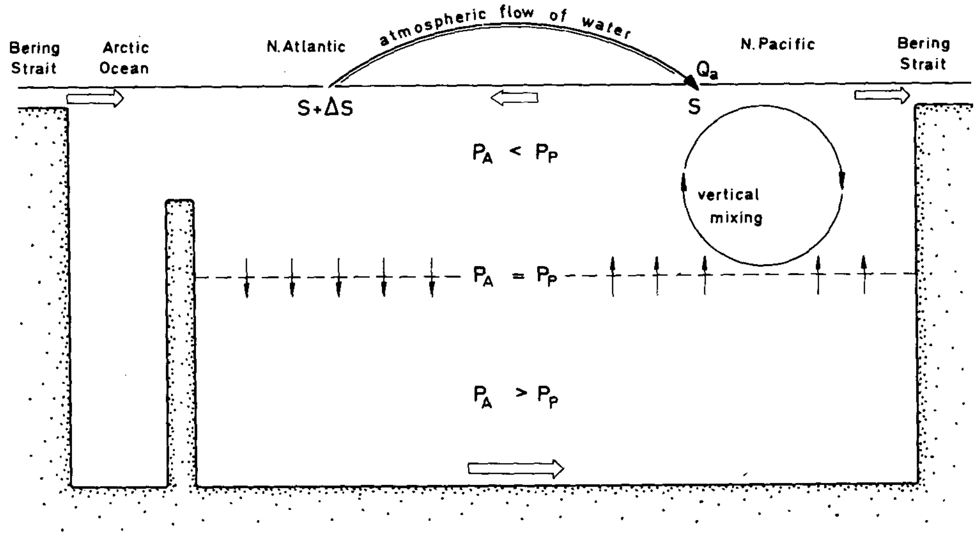

FIG. 1. Definition sketch showing the important features of the system. P_P and P_A are pressures in the North Pacific and North Atlantic respectively. The horizontal pressure gradient is zero where $P_P = P_A$.

2. An estimate of the mean sea level difference between the Bering Sea and the Arctic Ocean

In order to calculate the mean sea level difference between the Bering Sea and the Arctic Ocean, we need to know the vertical distribution of density down to the assumed level, at depth h_0, of the zero horizontal pressure gradient. We cannot use hydrographic stations in the Arctic Ocean because the sill depth between the Arctic and the Atlantic is only about 800 m. We therefore choose one hydrographic station in the Irminger Sea (*Carnegie* Station 10, 59°19′N, 34°15′W) and one in the Bering Sea (*Chelan* 1934, station 117, 54°12′N, 168°05′W), see Fig. 2. The vertical σ_t-distributions at these two stations are given in Table 1.

Using the densities in Table 1 we find that the sea level difference between the two stations $\Delta h'$ is ~65 cm. We have here assumed the horizontal pressure gradient to vanish at h_0 equal to 1100 m, cf. the concept of the depth of no horizontal motion (see e.g., Defant, 1960). Owing to the existence of the less dense polar surface water in the Arctic Ocean, the sea level there should be higher than in the Irminger Sea. This difference should be approximately 15 cm (a layer of thickness 200 m and salinity deficiency of about 1‰). Thus, the sea level difference between the Bering Sea and the Arctic Ocean Δh should be about 50 cm. Systematic sea level variations of a period of one year, e.g., due to seasonal heating and cooling of the sea surface, certainly occur; see e.g., Lisitzin (1974). However, the amplitudes and phases are approximately equal in the North Atlantic and the North Pacific. Thus, the sea level difference between the two oceans should be fairly constant throughout the year. This may not be true for coastal stations where a varying freshwater content in the surface layers may add to the yearly cycle of sea level variations.

The estimate of the sea-level difference is of course dependent upon the choice of the depth of the zero horizontal pressure gradient. However, a shift of this depth by a few hundred meters will give only a minor change. Since the sea level may vary on small horizontal-length scales as a result of geostrophic currents, the estimate of the sea level difference is also dependent upon the location of the hydrographic stations used for the calculation. The two stations used here are both considered to be situated in areas that are rather unaffected by currents. However, a better estimate of the sea level difference should be obtained by taking a properly weighted average over several hydrographic stations in the three regions. At the present level of modeling, such refinements are thought not to be worthwhile. Reid (1961) found, by averaging data from the whole North Pacific and North Atlantic respectively, that the North Pacific stands 41 cm higher than the North Atlantic relative to the 1000 db surface. He also made reference to the leveling surveys indicating that along the Pacific coast of the United States, the mean sea level is ~50 cm higher than along the Atlantic coast.

3. Flows through the Bering Strait

From the estimates above, the sea level drop from the Bering Sea to the Arctic Ocean seems to be ~0.5 m. The transport driven by this longitudinal sea surface slope in the Bering Strait was parameterized by Stigebrandt (1981b) within the frame of nonrotating channel hydraulics. Using a quadratic friction law and a channel of (constant) rectangular cross section, it

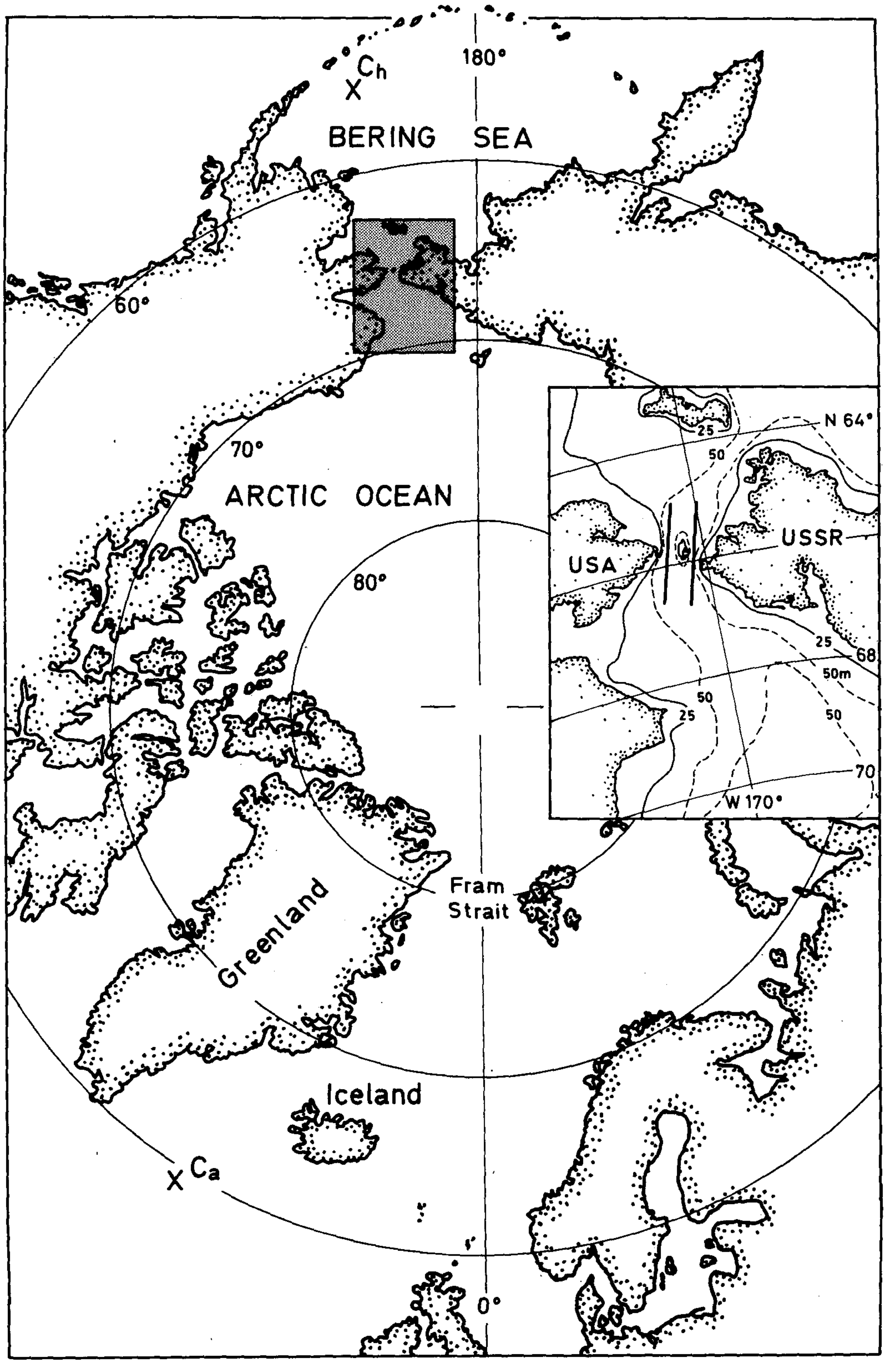

FIG. 2. Map showing the location of the two hydrographic stations (Ch = *Chelan* and Ca = *Carnegie*). The inset shows a map over the Bering Strait.

TABLE 1. The density at different depths (σ_t-values) in the Bering Sea (*Chelan*, Station 117, 24/8-1934) and in the Irminger Sea (*Carnegie*, Station 10, 30/7-1928).

Depth (m)	Chelan Station 117 (σ_t)	Carnegie Station 10 (σ_t)
0	25.51	26.77
10	25.55	26.82
25	25.76	26.82
50	26.24	26.92
75	26.42	27.29
100	26.54	27.50
150	26.65	27.59
200	26.74	27.62
300	26.84	27.64
400	27.00	27.64
500	27.07	27.64
600	27.15	27.69
800	27.30	27.74
1000	27.39	27.75

was found that the observed transport [about 1.5×10^6 m^3 s^{-1} (1.5 Sv)] was obtained for a model channel of width $W = 50$ km, depth $H = 50$ m and length $L = 200$ km if a "normal" value (0.003) of the drag coefficient c_d was used. In this class of channel hydraulics, the parameterization of the flow Q is (including end effects)

$$Q = WH\sqrt{g\Delta h}[1/(1 + c_d L/H)]^{1/2},$$

where g is the acceleration of gravity. However, one may raise a number of objections against the usage of this specific parameterization for the Bering Strait flow. The model strait and the real strait are not geometrically similar. The real strait has the width of the model strait over a much shorter distance than 200 km. There are, however, vast shallow areas on both sides of the main strait that certainly contribute to the total frictional losses. A general method to calculate the model L, which is a function of the topography of the Bering Strait, is not available.

Another objection against the parameterization above concerns the omission of the effects of the rotation of the earth. Such effects may enter in several ways. One way would be the vertical shrinking of the frictional bottom-boundary layer leading to a secondary, transversal circulation. However, the strait is so shallow that the bottom Ekman layer should extend to the sea surface. Another effect is connected with the "effective" value of Δh in a rotating system.

Further objections against the suggested parameterization concern frictional effects of an ice cover upon the flow, effects of superimposed barotropic fluctuations and effects of possible vertical stratification.

From the preceding discussion it follows that a good parameterization of the flow in the Bering Strait could have the following appearance

$$Q = A\sqrt{g\Delta h}(h_1 h_2), \qquad (1')$$

where A is the minimum cross-sectional area of the channel and the nondimensional functions h_1 and h_2 are expected to describe the dependence of Q upon effects of friction and the rotation of the earth respectively. Functional forms for h_1 and h_2 are not proposed in this paper. Here, it is sufficient to conclude that the transport through the Bering Strait is strongly related to the topography of the strait and to the sea level difference between the ends.

Coachmann *et al.* (1975) estimated, from summer current measurements in the Bering Strait, the transport to be slightly above 1.6 Sv. Later, Coachmann and Aagaard (1981) adjusted this transport estimate to below 1 Sv. The adjustment was based upon information obtained from more recent current measurements in a section north of the Bering Strait. The measurements also covered the earlier undocumented winter season.

The density difference between the North Atlantic and the North Pacific is mainly caused by differences in salinity. Earlier we introduced a level, at depth h_0, of zero horizontal pressure gradient. Above this level we can introduce a mean vertical salinity difference ΔS between the two oceans (the sea level difference is $\Delta h'$, see Section 2) by the following definition:

$$\rho\Delta h' = \rho_0\beta\Delta S h_0, \qquad (1)$$

where β is defined by the linear equation of state for sea water $\rho = \rho_0[1 + \beta(S - S_0)]$ and where (ρ_0, S_0) is a reference water. Using the approximation $\rho/\rho_0 \approx 1$, the definition equation for ΔS may be written

$$\Delta S = \frac{\Delta h'}{h_0\beta}. \qquad (2)$$

With $h_0 = 11 \times 10^4$ cm, $\Delta h' = 65$ cm and $\beta = 8 \times 10^{-4}$ (‰)$^{-1}$ one obtains $\Delta S = 0.75$‰. Taking the temperature effect into account, this result agrees quite well with that found by Reid (1961). He shows that, averaged over the whole ocean, the upper 1000 m of the North Atlantic is about 1‰ saltier and 2.2°C warmer than the North Pacific.

Neglecting the slight difference between Δh and $\Delta h'$, we find that $\Delta h \propto \Delta S$ and

$$Q = Af(\Delta S)h_1 h_2, \qquad (3)$$

where $f(\Delta S)$ is essentially proportional to the square root of ΔS. Earlier in this section we found good arguments for Q being strongly dependent upon Δh. If this is correct, the volume transport through the Bering Strait is determined by the large-scale salinity distribution in the world ocean.

The freshwater flux through the Bering Strait is also of great interest. We choose $S_A = 35$‰ as a reference salinity. This is the salinity of the Atlantic water flowing into the Arctic Ocean through the Fram Strait [see Aagaard and Greisman (1975)]. These authors also give the mean salinity of the water flowing into the

Arctic Ocean through the Bering Strait S_B to be 32.4‰. Thus, with Q in the range suggested above, the inflow of freshwater through the Bering Strait is $Q\Delta S_1/S_A \approx 0.1$ Sv where $\Delta S_1 = S_A - S_B$. This estimate of the freshwater flow through the Bering Strait agrees with that given by Hall and Bryden (1982). The barotropic current fluctuations u' may contribute to the freshwater flow through the strait if there is a correlation between u' and fluctuations in the salinity. To the best of the present author's knowledge, an estimate of the magnitude of this possible transport component has not yet been reported in the literature.

Here, ΔS_1 may be split into two components in the following manner: $\Delta S_1 = \Delta S + C$, where C is largely determined by regional and local conditions (cf. Warren, 1983). The freshwater flow through the Bering Strait may therefore be written

$$Q\Delta S_1/S_A \sim f(\Delta S)(\Delta S + C)/S_A.$$

If ΔS goes towards zero, the barotropic driving mechanism disappears $[f(\Delta S) \rightarrow 0]$. The freshwater flow then becomes zero even if C is large.

In order to get a better conception of the magnitude of the freshwater flow through the Bering Strait, we estimate that it corresponds to an annual excess precipitation and river runoff to the North Pacific of ~4 cm or an annual excess evaporation over the North Atlantic of ~8 cm.

4. The North Pacific: A global-scale estuary

We will now exploit the idea, put forward in the Introduction, that the North Pacific may be considered an estuary on the global scale. The atmospheric net flow of freshwater from the North Atlantic to the North Pacific is denoted by Q_a. We assume that the fraction α of Q_a returns to the North Atlantic through the Bering Strait (the rest passes the equator). According to Weyl (1968) the main atmospheric water transport from the Atlantic to the Pacific takes place across Central America. Deffeyes (cited by Weyl) estimated from atmospheric data this flow to be 0.1 Sv.

In the last section, we estimated the freshwater transport through the Bering Strait to be about 0.1 Sv. Thus, α may be close to 1. From this it follows that the net freshwater flow across the Pacific equator should be small (a conclusion that should be checked by using oceanic and atmospheric data). One explanation for this might be the Pacific equatorial surface currents which are primarily zonal and may prevent communication between the North and South Pacific.

In an ordinary small-scale estuary like a fjord, it is easy to trace the freshwater from the river to the estuary mouth. In an estuary with extremely small through-flow (normalized by the horizontal surface area of the estuary), like the Pacific, the freshwater turnover within the estuary is much larger than the through-flow (evaporation and precipitation are both about 1 m year^{-1}, precipitation plus runoff minus evaporation is 0.04 m year^{-1}) and it would not be easy to follow a labeled freshwater parcel coming from the Atlantic Ocean during its passage through the estuary. This is also evident from the very large residence time for freshwater in the Northern Pacific. This should be on the order of 10^3 years. [The upper "brackish" layer in the North Pacific contains approximately 2×10^6 km^3 of freshwater (relative to Atlantic water of salinity 35‰) and the through-flow of freshwater is approximately 3×10^3 km^3 year^{-1}.

For a steady state, where ΔS is constant, the freshwater balance for the upper layer in the North Pacific requires

$$Q \cdot (\Delta S + C)/S_A = \alpha \cdot Q_a, \tag{4}$$

or using Eq. (1),

$$f(\Delta S) \cdot (\Delta S + C) = (\alpha Q_a S_A)/(A \cdot h_1 \cdot h_2). \tag{5}$$

If the functions h_1 and h_2 were known, one could, for instance, calculate the permanent change of ΔS caused by a permanent change of Q_a. Eq. (5) clearly shows that changes of the topography of the Bering Strait, which will directly effect A and h_1 (for well-behaved changes, A and h_1 respond in the same direction), cause changes in ΔS. Thus, ΔS can be said to be topographically controlled by the Bering Strait.

During glaciations, the sea level in the ocean falls (and thereby A and h_1 decrease) because of the storage of water on the continents. Eq. (5) then shows that ΔS should increase during such circumstances. However, to the best of this author's knowledge, there is no empirical evidence for this as yet.

Here ΔS may be regarded as rather stable for time periods of at least a few hundred years (a fraction of the residence time for the "upper layer" in the North Pacific). The volume transport through the Bering Strait should thereby be stable on the same time scale. The freshwater transport through the Bering Strait, however, can possibly change on much smaller time scales as it also depends on the regionally and locally determined C. The possible regulation of the large Alaskan rivers emptying into the Pacific Ocean or regional climate changes in the northernmost part of the Pacific might well appreciably change C, and thereby the freshwater flux to the Atlantic Ocean and eventually (for more permanent changes of C) ΔS also.

If the water depth in the Bering Strait becomes very small or zero, the contemporary freshwater balance is, of course, no longer possible. In this case, the freshwater transported to the North Pacific by the atmosphere has to find a different path back to the North Atlantic. The accompanying salinity difference between the two oceans is determined by the dynamics of that flow pattern.

5. Discussion

Undoubtedly, the net atmospheric freshwater transports between the oceans are crucial for the large-scale hydrographic state of the world ocean and also, the vertical circulation and the formation of deep water in the ocean. The estimated salinity difference of $\sim 1‰$ between the North Atlantic and the North Pacific is, by the estuary analogy, supposed to be the salinity difference between the surface layer and the deep layer which is evidently large enough to prevent deep water formation in the North Pacific. The low surface salinity in the northern North Pacific (related to our C) further prevents the formation of deep water (see Warren, 1983). Furthermore, other factors, such as the existence or nonexistence of semi-enclosed seas, may be of importance (see Warren, 1981). However, this observational fact further strengthens the analogy between the vertical circulation in the North Pacific and that found in small-scale estuaries with a net freshwater supply. Deep water drawn into the basin from outside (the South Pacific) is slowly entrained into the surface layer. In a steady-state situation, the effects of vertical mixing and upwelling must balance so that surfaces of constant properties (e.g., salt, temperature, plant nutrients) are kept in steady positions.

A rough estimate of the residence time of the deep water in this gigantic estuary may be obtained by dividing the volume of deep water ($\sim 2 \times 10^8$ km^3) by the rate of upwelling ($\approx$ the flow rate through the Bering Strait $\approx 5 \times 10^4$ km^3 per year). This gives a residence time of 4000 years. The accompanying ocean-wide averaged upwelling velocity is on the order of 1 m year^{-1}. This estimate is in agreement with the estimate of the residence time for the upper layer ($\sim$1000 m thick) of approximately 1000 years stated earlier in this paper. The residence times for the lower and upper layers obtained above are believed to be upper bounds. Possible water exchanges with the South Pacific may lead to shorter residence times. Note that the residence time for the water in the upper layer has nothing to do with the "age" of the water (the time that has elapsed since the water last was in contact with the atmosphere). However, as there is no local convection reaching into the lower layer in the North Pacific, the mean age of the water in this layer is at least as great as half the residence time.

Our estimate of the upwelling velocity ($w \approx 1$ m year^{-1}) is low compared to earlier estimates. Craig (1969) and others have utilized one-dimensional advection–diffusion models. The value of the ratio k/w, where k is the (constant) vertical diffusion coefficient, is obtained from the vertical distribution of conservative tracers. The vertical distribution of one additional, decaying or nonconservative tracer is needed in order to determine k and w. Using carbon data, Craig estimated $w \approx 5$ m year^{-1} and $k \approx 1.6$ cm^2 s^{-1}. However, Craig's estimates may quite well be too high by a factor of 5 for the following reason.

The turbulent energy dissipation per unit volume is kN^2/Rf where N^2 is the buoyancy frequency and Rf is the flux Richardson number (Rf is the fraction of the turbulent energy used for work against the buoyancy forces). A reasonable mean value of N^2 in the interval 1000–4000 m is 2×10^{-6} s^{-2} and one often found value of Rf in geophysical systems is 0.05 (see e.g., Stigebrandt, 1976). This gives a dissipation rate of $\sim$20 erg s^{-1} in a vertical column with a unit horizontal cross-sectional area (1 cm^2), reaching from the depth 1000 m down to 4000 m. The wind is known to be the dominant energy source for motions in the sea. The energy transfer from the wind to the sea is $\boldsymbol{\tau} \cdot \mathbf{v}$ where $\boldsymbol{\tau}$ is the wind stress and $\mathbf{v}$ is the surface drift ($|v| \approx 0.03 \times W$, where W = wind speed). If we assume an ocean-averaged time-mean wind stress of 1 dyn cm^{-2} [$(\overline{W^2})^{1/2} \approx 8$ m s^{-1}, where the overbar denotes the averaging process], we obtain a time and space mean energy transfer of about 25 erg cm^{-2} s^{-1} through the sea surface (which probably is on the high side). Thus, if our mean wind stress estimate is reasonable, the high value of the vertical diffusion coefficient estimated by Craig and others means that almost all of the energy transferred from the wind into the water is dissipated in the depth interval 1000–4000 m! This, of course, cannot be the case; most of the energy supplied by the wind is certainly dissipated in the upper layers. From this we infer that the value of k is probably strongly overestimated in earlier estimates. (People seem to have been unconsciously governed by early estimates of the rate of deep water formation in the world ocean "giving" the value of w). If the one-dimensional advection–diffusion model holds, the ratio k/w is known; thus w must also have been strongly overestimated. In fact, the value of w estimated in this paper (implying $k \approx 0.3$ cm^2 s^{-1}) may be more in accord with energy constraints. The findings presented here may give a reasonable explanation for the great age of North Pacific deep waters (cf. Worthington, 1981).

Stigebrandt (1981a) showed that the existence of the pack ice in the Arctic Ocean is greatly secured by the contemporary large inflow of low-salinity water from the Pacific. However, a large reduction of this flow would make the Arctic pack ice more susceptible to disturbances and would promote the disappearance of the ice cover. Although the Arctic Ocean pack ice covers only 3% of the surface area of the Northern Hemisphere, it has been suggested (Donn and Ewing, see Weyl, 1968) that the disappearance of the ice cover may trigger a glaciation in the northern parts of this hemisphere (see Goody, 1980). Thus, the flow of freshwater through the Bering Strait (determined by the global ΔS and the regional C) may play an important role in climate dynamics. As the residence time for water parcels in the upper layer of polar surface water in the Arctic Ocean is only on the order of 10 years, large decreases of C on the same time scale might lead to drastic changes in the Arctic ice conditions (cf. Stigebrandt, 1981a).

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