# China's Development Could Lead to Bottom Water Formation in the Japan/East Sea

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## ABSTRACT

Using hydrographic data and box models, it is shown that the presently discussed diversion of rivers such as the Yellow or the Yangtze for agricultural use is likely to cause the renewal of Bottom Water formation in the Japan/East Sea. Such formation was common (near the Siberian coast) in the 1930s, 1940s, and 1950s, but subsided since that time due to a warming trend (accompanied by a decreased salinity due to the melting of ice). Since a diversion of freshwater is analogous to evaporation, a (diversion induced) increase of salinity is expected and the increase is large enough to allow Bottom Water formation even at the present-day cooling rates. Even a modest diversion of "merely" 3000 $m^3$ $s^{-1}$ (which is 10% of the total freshwater flux) will probably cause Bottom Water formation at a rate of roughly 750 000 $m^3$ $s^{-1}$. This is the first study that predicts anthropogenic *reversal* of an existing vertical structure in a semienclosed sea.

## 1. Introduction

The Japan/East Sea (JES) is a deep (> 3500 m) semienclosed basin fed by small amounts (approximately 0.1 Sv where 1 Sv ≡ $10^6$ $m^3$ $s^{-1}$) of relatively fresh and cold water from the (poorly observed) Sea of Okhotsk in the north (Fig. 1). This flow is enhanced by the melting of ice (Seung and Kim 1993; Holloway et al. 1995; Martin and Kawase 1998; Riser et al. 1999) in the western part of the JES. Superimposed on this fjordlike circulation is a much stronger throughflow of about 1½ Sv of warm and salty Pacific water entering the JES through the shallow Tsushima Strait (< 100 m) and exiting through the Tsugaru (< 130 m) and Soya Straits (< 70 m). This throughflow originates in the Kuroshio and the Taiwan Currents (Isobe 1999), both of which are primarily driven by the winds over the Pacific rather than the local winds directly above (Nof 2000). Although the Tsushima Current is salty

relative to the water entering the JES from the north, it is fresh compared to the Pacific water from which it originated because of significant dilution by freshwater from both the Yangtze River and the Yellow River (Tomczak and Godfrey 1994). For other general aspects of the JES circulation the reader is referred to Minato and Kimura (1980), Toba et al. (1982), Sekine (1986, 1988), Sugimoto (1990), Preller and Hogan (1998), and Hogan and Hurlburt (2000).

### a. Background

The JES is subject to outbreaks of very cold Siberian air during winter. During a typical January the temperature difference between the air and the water can reach more than 15°C (see, e.g., Seung and Yoon 1995), a difference that is twice as much as the 7°C difference between the air and the water in the convective region of the Gulf of Lion in the Mediterranean Sea (see, e.g., Marshall and Schott 1999). It has been argued that due to this JES cooling and the relatively salty water brought in by the Tsushima Current, deep water occupying the depth range of 200–2000 m is regularly formed (Gamo et al. 1986; Sudo 1986; Senjyu and Sudo 1994; Seung and Yoon 1995; Yoshikawa et al. 1999). [This water is sometimes referred to as the Japan Sea Proper Water and should be distinguished from the Bottom Water (BW), defined as water occupying depths greater than 2000 m. Note

Fig. 1. The JES and its neighboring bodies of water. Relatively fresh and cold water enters the JES from the north (through the Tatar Strait). Superimposed on this fjordlike circulation is a flow of Pacific water (entraining the freshwater from the Yangtze and Yellow Rivers), which enters through the Tsushima Strait and exits via the Tsugaru and Soya Straits. The JES is subject to strong winter cooling due to outbreaks of cold Siberian air off Vladivostok. A close-up of the squared region off Vladivostok is shown in Fig. 4.

(which used to carry approximately 3000 $m^3$ $s^{-1}$) and part of the Yangtze River (which carries as much as 30 000 $m^3$ $s^{-1}$) for agricultural use. For our calculations, we shall assume that the diversion will cause a loss of at least 3000 $m^3$ $s^{-1}$ of freshwater input to the ocean, an amount comparable to the known freshwater loss due to the construction of dams and agricultural development in other parts of the world (e.g., the Mediterranean Sea).

### b. Approach

We shall take the so-called box model approach were horizontal and vertical variations within each layer of the conceptual box are ignored. Opinions vary as to how useful such an approach is because horizontal variations can at times be important. One school of thought argues that detailed numerical models are always needed and that box models do not provide useful information. Supporters, on the other hand, state that when observations are scarce and when one does not know how to parameterize crucial numerical procedures (such as convection), box models are not only needed but are, in fact, preferable. Our opinion is that, with the present-day understanding of the JES, a box model approach is an appropriate *first stage* of looking at the problem.

### c. Presentation

Our analysis will be presented in three stages. First, using *T–S* diagrams and the box model approach, we shall show that the present situation in the JES is not very far from that corresponding to regular bottom water formation (section 2). Since surface water regularly sinks to the 500–1000-m level today, we will not be concerned with this surface water stability but rather focus our attention on the conditions requiring the sinking of water from 500 to 2000–3000 m. For this *T–S* presentation, we shall assume that the *mean* salinity of the JES will be raised by the diversion and that the local salinity in the convection-prone area of

also that Riser et al. (1999) argues that at times such deep water is not formed.]

Oxygen measurements suggest that Bottom Water (> 2000 m) was also regularly formed *until* the 1950s (Gamo et al. 1986; Nitani 1972; Minobe 1996; Varlamov et al. 1997); since that time a warming trend accompanied by a decrease in salinity has reduced the production of this BW to almost zero (Gamo et al. 1986). Due to the relatively shallow sills and the lack of new BW production, present-day BW does not freely communicate with the Pacific Ocean and, consequently, it is trapped within the JES (Fig. 2).

It is hard to tell how much freshwater will be diverted by the Chinese as a part of the regional development and the completion of the Three Gorges Dam because such information is not presently made public. The official aim of the Three Gorges Dam is flood control and the generation of hydroelectric power (neither of which should have an effect on the average freshwater flux to the ocean) but there have been statements regarding the diversion of the Yellow River

Vladivostock, Russia, will be raised by the same amount.

Second, we shall present a detailed comparison of the convective region in the Mediterranean Sea (MS) with that of the JES (section 3). We shall show that both seas possess all three requirements for deep convection: (i) both regions are subject to cyclonic circulation that elevates the isotherms and thus brings deep water closer to the surface; (ii) both regions are situated next to land where cold outbreaks of continental air take place in the winter; and (iii) in both regions the stratification beneath the surface layer is weak (most likely due to earlier convection). Based on this comparison, we shall argue that the JES is not very far from a state of deep convection. We shall further argue that, since it was subject to such convection in the not very distant past, it is very close to a state of deep convection in the present. So much so that a diversion of rivers is sufficient to cause it to become convective. (In this scenario, it is assumed, of course, that the convection is induced by cooling rather than brine rejection.)

It is useful to point out here that, in the familiar Mediterranean Sea case, dense water is formed by initial evaporation in the Levantine basin (analogous to our Yellow and Yangtze River diversion) and a subsequent atmospheric cooling south of France. For this well-known case it has already been demonstrated (Nof 1979; Béthoux et al. 1990; Rohling and Bryden 1992) that the construction of the High Aswan Dam and the diversion of a few thousand cubic meters per second had important consequences on the circulation. [In this context, the reader is also referred to Rahmstorf (1997) who performed numerical experiments with oceans subject to varying surface salinities and to Semtner (1984, 1987) who examined the sensitivity of the Arctic Ocean to the diversion of rivers. Another related case of interest is that of Steinhorn et al. (1979) who *observed* an anthropogenic turnover of the Dead Sea.]

Third (section 4), we shall present a box model and compute the new steady state that will be reached [in approximately 300 years, the residence time of the JES (Tomczak and Godfrey 1994)] after the diversion is completed. Before beginning our detailed presentation, it is pointed out again that the JES is not as well observed as the MS [primarily due to the very long inaccessibility of its western part to western ships (which was altered with the collapse of the USSR)]. Consequently, it is virtually impossible to accurately determine (using both observation and models) what a diversion of rivers would cause. As mentioned, our
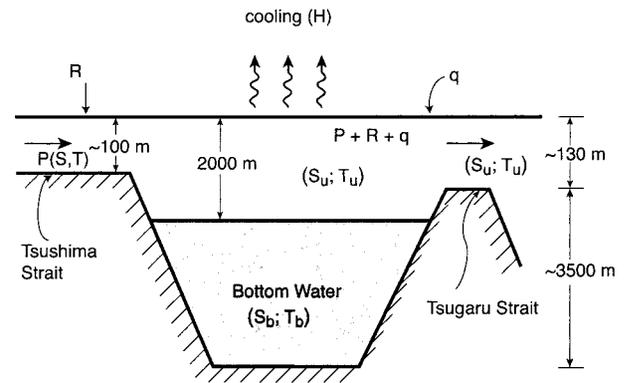


FIG. 2. Schematic diagram of the present-day equilibrium state in the JES. A volume flux $P$ [approximately 1½ Sv (1 Sv = $10^6$ m$^3$s$^{-1}$)] of Pacific water (with a salinity $S$ and temperature $T$) entrains freshwater $R$ (approximately 30 000 m$^3$ s$^{-1}$ of fresh Yangtze and Yellow River water) as it enters the JES. Within the JES, it is further diluted with small relatively freshwater flux q (~0.1 Sv) originating from the Sea of Okhotsk and its salinity becomes $S_u$. (The subscripts u and b indicate that the variable in question is associated with the upper and bottom water.) At the same time, it is being strongly cooled down from $T$ to $T_u$ by outbreaks of cold air from Siberia. Due to the relatively low salinity of the present-day water entering the sea, and the relatively high salinity of the BW presently trapped in the JES, the cooling is not sufficient to produce new BW, i.e., the density difference between the water within the sea's upper layer and the present BW is negative even during periods when strong cooling is taking place. Since there is no BW production, present BW does not spill out over the sill. It is argued that even with a 10% diversion upstream the cooling-induced density difference will become positive and deep convection will take place (see text and Fig. 3; "wiggly" arrows indicate atmospheric cooling).

box model approach and its back-of-the-envelope calculations are probably the best that one can do with the issue at this stage. Further measurements of the freshwater path in the JES and more detailed hand-in-hand modeling efforts that incorporate convective processes will most likely yield better predictions.

## 2. Old and new density structures

It is a simple matter to show that, with our (box model) assumption of a uniform salinity change in the upper layer, a diversion of rivers would cause BW formation. To see this, we examine Fig. 3 where the potential density contours as a function of temperature and salinity (relative to 3000 dbar) are shown. Superimposed on these contours are (the old and new) $T$–$S$ data analyzed by Nitani (1972; for the Liman Current region) and the data analyzed by Riser et al. (1999; for the northern JES). As Riser et al.'s analysis focuses
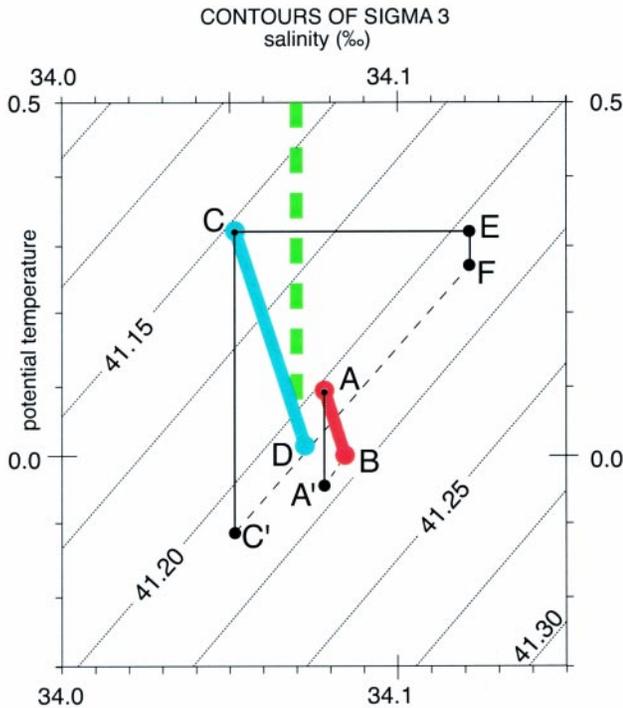
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salinity (‰)

FIG. 3. A $T$–$S$ diagram showing the historical density structure in the Liman Current according to Nitani (1972) and the present density structure in the northern JES according to Riser et al. (1999). The thick blue line (connecting C and D) and thick red line (connecting A and B) correspond to Nitani's (1972) analysis whereas the dashed green line corresponds to Riser's et al. (1999) data. The blue line shows a typical *no* BW formation era (1960–70) whereas the red line corresponds to a formation era (1940–50). Points A and C correspond to 500-m depth and points B and D to 2500–3500-m depth range. We see that during the formation era (red line), cooling of merely 0.15°C was sufficient to produce BW. To see this, note that water corresponding to point A had to be cooled an amount corresponding to AA′ in order to convect (i.e., in order for its density to be equal to that of the Bottom Water). On the other hand, during the no BW formation era (blue line), a much greater cooling of almost 0.5°C is required in order to produce deep convection. Namely, water corresponding to point C has to be cooled to a temperature corresponding to point C′ in order to produce BW. With a diversion of "merely" 3000 m³ s⁻¹ (corresponding to 10% of the Yangtze and Yellow River volume flux), the upper water salinity would increase by approximately 0.068‰ making the upper water considerably denser than it presently is (point E). This increase in density is sufficient to produce BW even with present-day cooling rates since it requires a cooling of merely 0.05°C, which is a third of the old cooling rate (see line EF).

on the present-day situation (with no BW production) whereas Nitani's data includes the old period of BW production, we shall focus our attention on Nitani's data. (Riser et al.'s data is presented here merely for a comparison.) Note that Nitani used all the available Japanese data since 1928. His plots contain perhaps a

hundred data points (from which the shown curves have been constructed). The data that he used were collected in both summer and winter but this should not matter for our analysis that addresses long-term processes.

Nitani's (1972) analysis displays two situations, a present-day no BW production era (blue line connecting C and D and corresponding to Nitani's data for 1960–70) and an old era (red line connecting A and B and corresponding to 1940–50) where BW production *did take place.* The upper points of these two curves (A and C) correspond to upper water at 500-m depth whereas the lower points (B and D) correspond to water in the 2500–3500-m depth range. (Unfortunately, there is no historical data for the surface water that is in direct contact with the atmosphere. Nevertheless, the analysis of the vertical density structure is still very useful as it will tell us how much cooling is needed to convect intermediate water already cooled to the density level present at the 500-m level.)

We see that, during the old era when BW was regularly produced (red curve), an intermediate water cooling of merely 0.15° C was sufficient to produce deep convection (see line AA′). During the no BW formation era (blue curve), on the other hand, a much stronger intermediate water cooling of approximately 0.5°C is needed to produce BW (see line CC′). Evidently, present-day cooling rates are just not sufficient to produce such a large temperature difference. Namely, $\alpha(T_u - T_b) + \beta(S_u - S_b) < 0$ where, as shown in Fig. 2, $T_u$ and $T_b$ are the no BW production era temperatures of the upper and bottom layers, $S_u$ and $S_b$ are the salinity of the upper and bottom layers, and $\alpha$ and $\beta$ are the expansion coefficients for temperature and salinity. Note that $S_u$ and $T_u$ correspond to point C whereas $S_b$ and $T_b$ correspond to point D (Fig. 3). (The subscripts "$u$" and "$b$" indicate that the variable in question is associated with the "upper" and "bottom" waters. For clarity, all variables are defined in both the text and in the appendix).

To see what happens when a fraction of the freshwater influx is diverted we first examine the conservation of salt flux into the JES. As shown in Fig. 2, we assume that an amount $P$ of Pacific water with a salinity $S$ and temperature $T$ approaches the Tsushima Strait. Without any river diversion, the salinity of the water entering the JES via the Tsushima Strait is $S[P/(P + R)]$, where $R$ is the (undiverted) freshwater mass flux. In other words, the salinity of the entering water is lower than the original Pacific salinity $S$ due to the dilution by the fresh river water. In the absence

of this dilution the salinity would increase by approximately $\Delta S = S \, \delta R/P$, where $\delta \, (< 1)$ is the fraction of the river flow that is diverted. For $\delta R = 3000 \ \mathrm{m^3 \ s^{-1}}$ (i.e., $\delta = 0.1$), $P = 1\,500\,000 \ \mathrm{m^3 \ s^{-1}}$ and an average salinity of 34‰ we find that $\Delta S = 0.068$‰. Increasing the surface water salinity (point C in Fig. 3) by this amount (without affecting its temperature) brings us from point C to point E (Fig. 3). Water with these diversion-affected properties now requires a cooling of merely 0.05°C in order to sink and form new BW (see line EF). Such a cooling is smaller than that associated with the BW formation era (0.15°C) leading us to the conclusion that BW will almost certainly be formed if a diversion of at least 10% is completed. Note that a complete diversion ($\delta = 1$) gives an incredibly high salinity increase of approximately 0.7‰, which would cause an immediate deep convection even without any cooling. In the next section, we shall show that, with the diversion completed, the JES will become essentially analogous to the MS.

## 3. Comparison of the convection in the Mediterranean Sea to that in the Japan/East Sea

The first convection directly observed in the Gulf of Lion was the 1969 convection (see e.g., MEDOC Group 1970; Gascard 1978; Schott and Leaman 1991). It helped identify the three now-familiar subprocesses active during the convection: "preconditioning" [where the cyclonic motion, O(100 km), brings weakly stratified fluid close to the surface], "deep convection" [where small-scale plumes of O(1 km) help break the Taylor–Proudman depth-independent constraint and cause high vertical velocities O(10 cm s$^{-1}$)], and "lateral exchanges" (where the convection is connected to the ambient fluid through processes on the scale of tens of kilometers). The convection in the Gulf of Lion is caused primarily by the mistral winds that bring cold dry air with a temperature of only 5°C to the relatively warm Mediterranean water (12°C). Although the average heat flux from the northwestern Mediterranean to the atmosphere during December, January, and February is no more than about 300 W m$^{-2}$, values as high as 100 W m$^{-2}$ were calculated for convective events (Leaman and Schott 1991). This strong cooling affects waters with an already enhanced salinity (due to the evaporation of about 100 000 m$^3$ s$^{-1}$ in the eastern Mediterranean) so that deep convection occurs.

The JES is subject to conditions very similar to those of the MS (Fig. 4). First, the convection regions in the two seas are approximately at the same latitude. Second, both regions have cyclonic circulation that brings deep water closer to the surface. Third, outbreaks of cold air occur in both regions. In the MS, the mistral brings in continental air 7°C colder than the surface seawater. In the JES, outbreaks of Siberian winds bring in masses of air 15°C colder than the surface water. The much larger air–sea surface temperature difference in the JES case reflects the fact that the JES is situated next to the eastern boundary of the continent whereas the MS is situated next to the western boundary of the continent (tempered by the Atlantic Ocean and the Gulf Stream).

It is difficult to compare the heat fluxes from the two seas because, as mentioned, the JES has not been observed to the extent that the MS has been. On the basis of the much greater air–sea temperature differ-
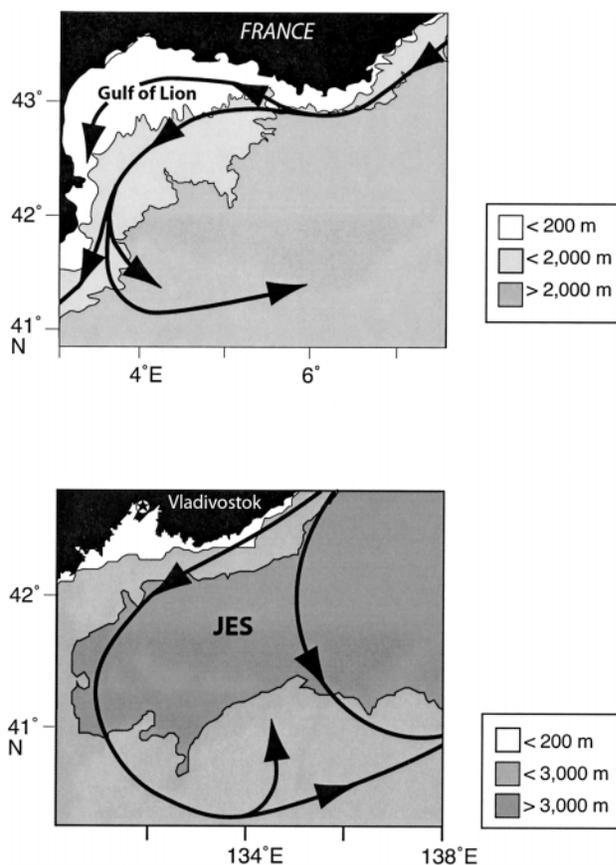


Fig. 4. The surface circulation in the northwestern Mediterranean (top, adapted from Marshall and Schott 1999) and the western JES (bottom, adapted from Tomczak and Godfrey 1994). Note that both areas are approximately at the same altitude and that both have cyclonic circulation.

ence in the JES case and the proximity of the JES to Siberia, it is expected that the JES heat flux would be greater than that of the MS. However, using recently released USSR observations and other data, Hirose et al. (1994) estimated the December heat flux in the convective region to be about 200 W m$^{-2}$, an amount comparable to that of the MS (300 W m$^{-2}$). For a complete discussion of the JES heat flux to the atmosphere see also Kato and Asai (1983), Hirose et al. (1996), Takata et al. (1997), and Na et al. (1999).

An important difference between the MS and the JES is the much smaller evaporation rate in the JES case (roughly 0.2 m yr$^{-1}$ compared to 1.0 m yr$^{-1}$). This probably explains why the BW formation process has been absent (from the JES) in the last forty years. As mentioned earlier, however, it is expected that a partial diversion of rivers will play a role analogous to that of evaporation in the MS. On the basis of our earlier analysis, we expect that this will be the case even though the diverted amount (a few thousand cubic meters per second) is much smaller than the apparent difference in the evaporation between the two seas.

## 4. The final equilibrium state

To examine our conclusion (regarding a 10% diversion) more closely and determine the final adjusted rate of BW production we consider the box model shown in Fig. 5. In this figure we schematically show the equilibrium state that will be reached after all the adjustments have been completed; that is, we show the structure that will be obtained after the diversion is completed and a new steady state is reached. We expect that two adjustments will be present during the transition to steadiness.

The first is the one associated with the arrival of the less-diluted Pacific water to the BW formation site (off Vladivostok). It is expected that this would take from a few months to many years from the time that the diversion is completed depending on the (very difficult to estimate) rate of horizontal mixing (due to either advection or eddy diffusion). Taking the horizontal scale as 500 km and the horizontal eddy diffusivity to be as low as $10^2$ m$^2$ s$^{-1}$, one arrives at a timescale of 70 yr, which should be regarded as an upper bound for the arrival of the new salinity. Once this happens, a new BW formation process will begin immediately. To accommodate the newly formed BW that continuously pours into the abyssal JES, the interface bounding the BW from the top will have to rise

until the BW will be able to continuously spill out over the Tsugaru Strait sill. Once this happens the old BW will be flushed out and, ultimately, a new equilibrium state will be established.

As in other similar situations (e.g., the Mediterranean or the Red Sea), it is expected that this new equilibrium state will be accomplished by a "hydraulic control" at the Tsugaru Strait [implying that the transport through the strait will be maximized (see, e.g., Smeed 1997; Nof 1979)]. Note that, without the wind-driven throughflow of Pacific water $P$ (which is superimposed on the cooling-induced flow), the JES would have had a two-way flow at the Tsugaru Strait and the picture would have been very similar to the familiar two-layer exchange between the Mediterranean Sea and the Atlantic Ocean. The fact that the total flow is expected to be one-directional rather than two-directional should have no bearing on the condition of hydraulic control as the condition involves the square of the velocity (and not its direction). However, because of the JES throughflow, the analysis below is not identical to that of the MS. In what follows, we shall use the condition of hydraulic control and the conservation of heat and salt to find the new equilibrium state.

### a. Hydraulic control

The control at the Tsugaru Strait corresponds to a Froude number of unity that can be expressed as
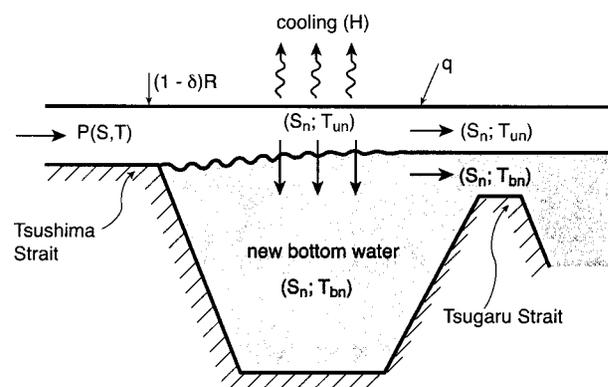


FIG. 5. A diagram of the equilibrium state after a diversion of $\delta R$ is completed and an adjustment period of approximately 300 yr (the flushing time of the JES) has passed. In this new state, the "old" BW (i.e., present-day BW with a relatively high salinity and temperature of $S_b$ and $T_b$, respectively) has been flushed out and new BW with a uniform salinity of $S_n$ and temperature $T_{bn}$ is present. The subscript n indicates that the variable in question is associated with the new state. It is assumed here that the effect of the relatively freshwater entering the JES from the Sea of Okhotsk ($q$) on the new deep convection is small and negligible (see text).

$$\frac{[P+(1-\delta)R]^2}{n} + \frac{[P+(1-\delta)R]^2}{1-n} = g * b^2 D^3, \quad (1)$$

where $P$ is the (known) no river-dilution Pacific volume flux, $\delta R$ the much smaller (known) diverted freshwater flux, $n$ the (yet unknown) fraction of the water column occupied by the upper layer at the Tsugaru Strait, $b$ the (known) average width of the strait ($\sim$17 km), $D$ its (known) average depth ($\sim$130 m), and $g*$ is the (yet unknown) "reduced gravity" given by

$$g* = \frac{g}{\rho_0}\alpha(T_{un} - T_{bn}).$$

Here, $\rho_0$ is a reference density, $T_{un}$ and $T_{bn}$ are the new upper and BW temperatures downstream from the cooling region (i.e., at the Tsugaru Strait), and, as before, $\alpha$ is the expansion coefficient for temperature. Note that the subscript "n" indicates "new," that is, the new state after the diversion is complete and a new steady state is reached. As before, the subscripts u and b indicate upper and bottom.

The low salinity water flux $q$ from the Sea of Okhotsk into the JES is neglected here on the ground that, according to recent estimates (Martin and Kawase 1998), it is small (0.1 Sv) compared to $P$, and on the ground that whatever enters the JES from the north is not going to be altered by the diversion in the south so it can simply be superimposed on our independent solution. Furthermore, in line with the box model approach, the new salinity of the convective JES is taken to be independent of depth (so that it does not affect the new density difference between the layers).

### b. Conservation of heat and salt

The system is also governed by the conservation of heat equation

$$P\,C_p[T - n\,T_{un} - (1-n)T_{bn}] = H, \quad (2)$$

where $H$ is the (known) heat loss to the atmosphere, $C_p$ is the heat capacity of water, and it has been taken into account that $R \ll P$. Finally, the conservation of salt implies

$$S_n = S\left[\frac{P}{P+(1-\delta)R}\right], \quad (3)$$

where $S_n$ is the new (uniform) salinity created by the diversion of freshwater ($\delta R$).

### c. Solution

It is easy to show that the function $n(1-n)$ [which is a part of relation (1)] reaches a maximum with $n = 1/2$ implying that maximum transport through the Tsugaru Strait will take place when both the upper and lower layers occupy half of the total depth $D$. As is done in other similar (hydraulically controlled) situations (e.g., Nof 1979), it is assumed here that the transport will indeed be maximized so that we can take $n = 1/2$. Together with R $\ll$ P, (1) and (2) then take the much simpler form,

$$4P^2 = \frac{g}{\rho_0}\alpha(T_{un} - T_{bn})b^2 D^3 \quad (4)$$

and

$$\left(T - \frac{T_{un} + T_{bn}}{2}\right) = \frac{H}{PC_p}, \quad (5)$$

which are two equations with two unknowns, $T_{un}$ and $T_{bn}$.

It is important to realize that, in this box model scenario, the maximization of the transport and the resulting $n = 1/2$ implies that half of the incoming surface water flux will become BW flux, that is, the BW production rate is $P/2$. For $P = 1.5$ Sv, $b = 17$ km, $D = 130$ m, and $\alpha = 1800 \times 10^{-7} k^{-1}$, (4) gives the reasonable value of 10°C for the mean temperature difference between the exiting upper water ($T_{un}$) and the exiting BW ($T_{bn}$). Similarly, an average heat loss of approximately 50 W m$^{-2}$, which is typical for the JES (see, e.g., Hirose et al. 1994, 1996; and Oberhuber 1988; but note that Na et al. 1999 argued that the average heat loss is even higher, about 100 W m$^{-2}$), an area of approximately 800 000 km$^2$ gives, with the aid of (5), a mean temperature reduction $[T - (T_{un} + T_{bn})/2]$ of 5°C. For a reasonable initial temperature (of incoming surface Pacific water) of $T = 10$°C we get an exiting surface water with an identical temperature $T_{un}$ of 10°C and a reasonable, much lower BW temperature of $T_{bn} = 0$°C.

The reader should be warned in this stage that, although the conclusion that new BW formation will take place if a diversion is completed is straightforward and fairly certain, the box model calculation (giv-

ing a BW production rate of approximately 0.75 Sv with a temperature of 0°C) has a fairly broad margin of error.

## 5. Summary

Prior to summarizing our results, it is appropriate to list again the limitations of our approach. First, the box model is formulated under the assumption that the diversion is complete in the sense that no diverted water reaches the JES via evaporation over the land and a subsequent precipitation over the sea nor via groundwater. In reality, some diverted water may eventually reach the sea and this may alter our results. Second, the box model approach assumes that both the upper and lower layers are well mixed both horizontally and vertically. Again, in reality, this is not the case as there are horizontal density gradients as well as vertical gradients within each layer. Other processes that may affect the results are the influence of the complex geometry of the Tsushima and Tsugaru Straits, which have been neglected. Also, we have not addressed the issue of the upper 500-m stability on the ground that convection of surface water above this level occurs even at the present-day salinity levels.

Despite all of these weaknesses, box models of the kind used here have been proven useful in numerous situations. There are many examples but the most closely related one is that of Nof (1979), which, using a similar model, predicted salinity alterations (due to the construction of the Aswan High Dam) 10 yr before they were actually observed (Béthoux et al. 1990; Rohling and Bryden 1992). In the present particular case of the JES, the choice of a box model is particularly sensible as we do not know the JES structure well enough to (numerically) model the diversion-induced convection in detail.

Our box model predicts that the diversion of rivers for agricultural use can have important effects on the meridional circulation in the Pacific Ocean due to the sensitivity of the JES to the surface salinity distribution. How the Pacific will respond to the formation of 0.75 Sv of BW in the JES is not clear. This amount (0.75 Sv) is similar to that of the MS (1 Sv) but, given that the Pacific is almost three times as broad as the Atlantic, its effect will probably be smaller than that of the MS. A detailed study analogous to that of Rahmstorf (1997) is needed in order to answer this question. Despite the expected moderate effect on the Pacific, it is expected that the response of the JES to even a partial diversion of the Yellow and Yangtze Rivers will be much more dramatic than that of the Mediterranean due to the construction of the Aswan High Dam. We expect to witness far-reaching environmental effects due to the ability of the system to flip from the present no BW production equilibrium state to a new, very different state where BW is regularly formed.

## Appendix: List of Symbols

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| $b$ | average width of the Tsugaru Strait |
| $C_p$ | heat capacity of water |
| $D$ | average depth of the Tsugaru Strait |
| $g^*$ | reduced gravity $\dfrac{g}{\rho_0}\,\alpha\,(T_{un}-T_{bn})$ |
| $H$ | heat loss to the atmosphere |
| $n$ | fraction of the water column occupied by the upper layer at the Tsugaru Strait |
| $P$ | volume flux of Pacific water approaching the Tsushima Strait |
| $q$ | low salinity water flux through the Tatar Strait |
| $R$ | freshwater influx from the Yangtze and Yellow Rivers |
| $S$ | salinity of incoming Pacific water |
| $S_n$ | new (uniform) salinity created by the diversion of freshwater $(\delta R)$ |
| $S_u, S_b$ | salinity of the upper and bottom layers |
| $T_{bn}$ | new bottom water temperature downstream from the cooling region |

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|---|---|
| $T_{un}$ | new upper water temperature downstream from the cooling region |
| $\alpha$ | expansion coefficient for temperature |
| $\beta$ | expansion coefficient for salinity |
| $\delta$ | fraction of diverted water ($\delta < 1$) |
| $\delta R$ | diverted freshwater flux |
| $\rho_0$ | reference density |

# References

Béthoux, J. P., B. Gentili, J. Raunet, and D. Tailliez, 1990: Warming trend in the western Mediterranean deep water. *Nature,* **347,** 660–662.

Gascard, J.-C., 1978: Mediterranean deep water formation, baroclinic eddies and ocean eddies. *Oceanol. Acta.,* **1,** 313–315.

Gamo, T., Y. Nozaki, H. Sakai, T. Nakai, and Y. Tsubota, 1986: Spacial and temporal variations of water characteristics in the Japan Sea bottom layer. *J. Mar. Res.,* **44,** 781–793.

Hirose, N., C.-H. Kim, and J.-H. Yoon, 1994: Heat budget in the Japan Sea. *Proc. 3rd CREAMS Workshop,* Fukuoka, Japan, Research Institute of Oceanography, 49–54.

——, ——, and ——, 1996: Heat budget in the Japan Sea. *J. Ocean. Soc. Japan,* **52,** 553–574.

Hogan, P. J., and H. Hurlburt, 2000: Impact of upper ocean–topographical coupling and isopycnal outcropping in Japan/East Sea models with 1/8° to 1/64° resolution. *J. Phys. Oceanogr.,* **30,** 2535–2561.

Holloway, G., T. Sou, and M. Eby, 1995: Dynamics of circulation of the Japan Sea. *J. Mar. Res.,* **53,** 539–569.

Isobe, A., 1999: On the origin of the Tsushima Warm Current and its seasonality. *Contin. Shelf. Res.,* **19,** 117–133.

Kato, K., and T. Asai, 1983: Seasonal variations of heat budgets in both the atmosphere and the sea in the Japan Sea area. *J. Meteor. Soc. Japan,* **61,** 222–238.

Leaman, K. D., and F. Schott, 1991: Hydrographic structure of the convection regime in the Golfe du Lion. *J. Phys. Oceanogr.,* **21,** 575–598.

Marshall, J., and F. Schott, 1999: Open-ocean convection: Observations, theory, and models. *Rev. Geophys.,* **37,** 1–64.

Martin, S., and M. Kawase, 1998: The southern flux of sea ice in the Tatarskiy Strait, Japan Sea and the generation of the Liman Current. *J. Mar. Res.,* **56,** 141–155.

MEDOC Group, 1970: Observations of formation of deep-water in the Mediterranean Sea, 1969. *Nature,* **227,** 1037–1040.

Minato, S., and R. Kimura, 1980; Volume transport of the western boundary current penetrating into a marginal sea. *J. Ocean Soc. Japan,* **36,** 185–195.

Minobe, S., 1996: Interdecadal temperature variation of deep water in the Japan Sea (East Sea). *Proc. 4th CREAMS Workshop,* Vladivostok, Russia, Research Institute of Oceanography, 81–88.

Na, J., J. Seo, and H.-H. Lie, 1999: Annual and seasonal variations of the sea surface heat fluxes in the East Asian Marginal Seas. *J. Oceanogr.,* **55,** 257–270.

Nitani, H., 1972: On the deep and bottom waters in the Japan Sea. *Research in Hydrography and Oceanography,* D. Shoji, Ed., Hydrographic Department of Japan Maritime Safety Agency, 151–201.

Nof, D., 1979: On man-induced variations in the circulation of the Mediterranean Sea. *Tellus,* **31,** 558–564.

——, 2000: Why much of the circulation in the Atlantic enters the Caribbean Sea and very little of the Pacific circulation enters the Sea of Japan. *Progress in Oceanography,* Vol. 45, Pergamon, 39–67.

Oberhuber, J. M., 1988: The Budget of Heat, Buoyancy and Turbulent Kinetic Energy at the Surface of the Global Ocean. Max Planck Institut für Meteorologie, March, Rep. 15, 19 pp.

Preller, R. H., and P. J. Hogan, 1998: Oceanography of the Sea of Okhotsk and the Japan/East Sea. *The Sea,* A. R. Robinson and K. H. Brink, Eds., John Wiley and Sons, Inc., 429–481.

Rahmstorf, S., 1997: Risk of sea-change in the Atlantic. *Nature,* **388,** 825–826.

Riser, S., M. Warner, and G. Yurasov, 1999: Circulation and mixing of water masses of Tatar Strait and northwestern boundary region of the Japan Sea. *J. Oceanogr.,* **55,** 133–156.

Rohling, E. J., and H. Bryden, 1992: Man-induced salinity and temperature increases in western Mediterranean deep water. *J. Geophys. Res.,* **97,** 11 191–11 198.

Schott, F., and K. D. Leaman, 1991: Observations with moored acoustic Doppler current profilers in the convection regime in the Golfe du Lion. *J. Phys. Oceanogr.,* **21,** 558–574.

Sekine, Y., 1986: Wind-driven circulation in the Japan Sea and its influence on the branching of the Tsushima Current. *Progress in Oceanography,* Vol. 17, Pergamon, 297–312.

——, 1988: On the seasonal variation in in- and outflow volume transport of the Japan Sea. *Progress in Oceanography,* Vol. 21, Pergamon, 269–279.

Semtner, A. J., Jr., 1984: The climatic response of the Arctic Ocean to Soviet river diversions. *Climatic Change,* **6,** 109–130.

——, 1987: A numerical study of sea ice and ocean circulation in the Arctic. *J. Phys. Oceanogr.,* **17,** 1077–1099.

Senjyu, T., and H. Sudo, 1994: The upper portion of the Japan Sea proper water; its source and circulation as deduced from isopycnal analysis. *J. Oceanogr. Soc. Japan,* **50,** 663–690.

Seung, Y.-H., and K. Kim, 1993: A numerical modeling of East Sea Circulation. *J. Oceanol. Soc. Korea,* **28,** 292–304.

——, and J.-H. Yoon, 1995: Some features of winter convection in the Japan Sea. *J. Oceanogr.,* **51,** 61–73.

Smeed, D., 1997: Seasonal variation of the flow in the strait of Bab al Mandab. *Oceanol. Acta,* **20,** 773–781.

Steinhorn, I., and Coauthors, 1979: The Dead Sea: Deepening of the mixolimnion signifies the overture to overturn of the water column. *Science,* **206,** 55–57.

Sudo, H., 1986: A note on the Japan Sea Proper Water. *Progress in Oceanography,* Vol. 17, Pergamon, 313–336.

Sugimoto, T., 1990: A review of recent physical investigations on the straits around the Japanese Islands. *The Physical Oceanography of Sea Straits,* L. Pratt, Ed., Kluwer Academic, 191–209.

Takata, T., S. M. Varlamov, N. Hirose, and J.-H. Yoon, 1997: Interannual variation of the surface heat flux in the Japan Sea. *Proc. 2nd CREAMS Int. Symp.,* Fukuoka, Japan, Research Institute for Applied Mathematics, 153–156.

Toba, Y., K. Tomizawa, Y. Kurasawa, and K. Hanawa, 1982: Seasonal and year-to-year variability of the Tsushima-Tsugaru Warm Current system with its possible cause. *La mer, 20,* 41–51.

Tomczak, M., and J. S. Godfrey, 1994: *Regional Oceanography: An Introduction.* Pergamon Press, 422 pp.

Varlamov, S. M., N. A. Dashko, and Y.-S. Kim, 1997: Climate change in the Far East and Japan Sea area for the last 50 years. *Proc. CREAMS '97 Int. Symp.,* Fukuoka, Japan, Research Institute for Applied Mathematics, 163–166.

Yoshikawa, Y., T. Awaji, and K. Akitomo, 1999: Formation and circulation processes of intermediate water in the Japan Sea. *J. Phys. Oceanogr., 29,* 1701–1722.