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Tritium and $^3$He in the Sargasso Sea

by William J. Jenkins

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ABSTRACT

The systematics of tritium ($^3$H), $^3$He and $^3$H-$^3$He dating are investigated for oceanic mixing systems responding to the North Atlantic surface water tritium transient. Although the $^3$H-$^3$He age is a single valued function of the “true” mixing age, verbatim acceptance of the $^3$H-$^3$He age will result in a substantial underestimate of the mixing age for systems with timescales approaching the time elapsed since the tritium transient (1964-1965) or greater. $^3$H-$^3$He data are presented for a number of stations taken recently in the Sargasso Sea and discussed qualitatively. Most remarkable are the $^3$H and $^3$He maxima associated with subtropical mode water and the density stratum usually associated with Mediterranean water. Preliminary evidence indicates that this latter water type is not necessarily of direct Mediterranean water origin, and suggests that it is a result of eddy exchange between the Sargasso Sea anticyclonic circulation at this stratum and higher tritium northern waters. The equivalent transport is of the order of 2.5 Sv.

It is demonstrated that the penetration of tritium into and generation of $^3$He within the subtropical North Atlantic main thermocline cannot be explained on the basis of a purely vertical transport model. Comparison of the vertical diagnostic model results for tritium with $^3$He flux balance calculations and other estimates of the vertical diffusivity suggests an upper limit of 0.1 cm$^2$/sec for vertical diffusivity and that vertical processes contribute less than ten percent to the mass balance of the thermocline. A simple, lateral mixing model is proposed which satisfactorily explains both the $^3$H and $^3$He distributions and their evolution in time. Although the exclusion of vertical diffusion as the major transport process rules out a number of thermocline models, this simple model is consistent with thermocline models incorporating isopycnal transport. A characteristic ventilation time for the main thermocline in the Sargasso Sea is estimated to be of the order of 50 y, while the oxygen minimum layer is characterized by a ventilation time in excess of seventy years.

Oxygen utilization rates are obtained for the upper 2 km, and are seen to decrease from a near-surface value of 0.5 ml/l/y to .004 ml/l/y at 1800 m, with an e-folding depth of the order of 350 m.

1. Introduction

It is difficult, if not impossible, to duplicate the oceanic environment sufficiently well to reliably measure the rates of oceanic chemical reactions in the laboratory. In a sense, this “Heisenberg Principle” of oceanography results from the fact that

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the very act of sampling or containment perturbs the system being studied, or that the rates of reaction are too slow to be perceptible on reasonable time scales. Furthermore, any attempts to quantify the \textit{in situ} processes by observation of oceanic distributions yield only qualitative and ambiguous estimates because of our ignorance of the magnitude and indeed the nature of the physical processes which affect or control the chemical distributions.

Yet an understanding of the rates and mechanisms of chemical and physical processes in the oceans is vital. The fact that the former is inextricably linked to the latter makes the task of acquiring this understanding more difficult. Nonetheless, as man's impact on the oceanic environment becomes significant, we must have quantitative knowledge of these processes in order to make rational policy decisions with regard to environmental protection. Further, the oceans represent a vast resource, the exploitation and utilization of which requires a detailed understanding of the ocean as a chemical and physical system. This "understanding" must go beyond the construction of mathematical analogs which mimic the observed oceanic distributions of chemical and physical properties. We must understand the principles and magnitudes of the exact processes and create a \textit{predictive} model of the system.

The aim of this paper is to use the observed oceanic distributions of two anthropogenic tracers (bomb-produced tritium and its daughter $^3\text{He}$) to discuss the relative roles of isopycnal and cross-isopycnal transport in the main thermocline of a subtropical gyre (the Sargasso Sea). First, the systematics of the tracers and $^3\text{H}-^3\text{He}$ dating are discussed (Sections 2 and 3), followed by qualitative discussion of the observed tritium and $^3\text{He}$ distributions (Section 4). In Section 5, an attempt is made to demonstrate that a general vertical advection-diffusion model even with variable diffusivity cannot satisfactorily explain both the observed tritium and $^3\text{He}$ profiles, and the results are discussed in light of other measurements and interpretations. The sensitivity of the $^3\text{He}$ distribution to the presence of some vertical diffusivity is tested, and a limiting value (upper bound) is placed on the vertical diffusion coefficient in the main thermocline. In Section 6, an alternate, isopycnal model is proposed which not only satisfactorily explains both the present tritium and $^3\text{He}$ profiles, but also the past available tritium profiles. The isopycnal model is then used (Section 7) to compute apparent oxygen utilization rates, which are compared with previous estimates.

2. Background

For some substances, mankind has already had a significant impact on the oceans. These substances invade the oceans through much the same pathways as future pollutants and for some of these, the source functions and chemistry are known well enough to allow their use as tracers of the transport processes by which they enter the oceans and are redistributed. These "Transient Tracers" (their distributions are
transient in nature, rather than in a steady state), represent a unique opportunity to study the dynamics of transport processes.

In many respects, tritium ($^3$H, the heaviest isotope of hydrogen) fulfills these requirements. Its distribution is in a transient state: the natural global tritium inventory (about 3 kg) was dwarfed by the massive (several hundred kg; Eriksson, 1965) production of tritium by nuclear weapons testing. This tritium entered the surface oceans in a strongly latitude and time-dependent pattern (see Bainbridge, 1963; Eriksson, 1965; Dreisigacker and Roether, 1978; Weiss and Roether, 1979). The source function for this substance has been moderately well characterized, although improvements must be made if refinements in modelling are to be accomplished. The "chemistry" of tritium is known too: virtually all tritium has entered the oceans as HTO, i.e., water, and in this respect, represents an ideal tag for water motion. Tritium is radioactive, decaying with a half-life of 12.26 y, which allows it to be measured in its low oceanic concentrations. Typical oceanic levels are of the order of 1 T.U. (1 T.U. = 1 Tritium Unit = 1 $^3$H atom per $10^{18}$ H atoms). By itself, however, anthropogenic tritium is used more as a "dye" tracer than a "clock" tracer, since its distribution is generally influenced more by dilution than by radioactive decay; although the latter, of course, must be accounted for.

Using a combination of oceanic tritium observations, tritium in precipitation data, observed $^3$H/$^{90}$Sr ratios and a larger $^{90}$Sr data set (relative to tritium), Dreisigacker and Roether (1978) have reconstructed mean North Atlantic surface water tritium concentrations as a function of time (Fig. 1: The Dreisigacker-Roether Function has been extended beyond 1975 as a simple exponential decay with a mean life of 13.4 y. This extrapolation fits current mid-latitude Sargasso Sea surface water concentrations from 1974-1977 within errors of measurement for this laboratory). The major concerns about this construct (the DRF) are that (a) it represents a mean response while the true response may be spatially variable due to latitudinal gradation in input and oceanic circulation, and (b) the results are, at least for the early part of the tritium transient, not tightly constrained by actual oceanic tritium data. In defense of the former, the gross latitudinal effect tends to be minimized due to the fact that at lower latitudes where the loading of tritium was less, the mixed layer tends to be thinner, leaving the concentration of tritium roughly ($\pm 10\%$) the same (Dreisigacker and Roether, 1978; Weiss and Roether, 1979). Further, vigorous recirculation within the subtropical gyre serves to homogenize the surface layers. With regard to the latter, it should be possible with future data to further constrain at least the surface processes and reconstruct the source function in a somewhat more sophisticated fashion. At this point in time, the Dreisigacker and Roether "source function" represents a valuable starting point to modelling the penetration of tritium into the deep North Atlantic.

The daughter product of tritium is $^3$He, a stable, inert gas. $^3$He is the less abundant isotope of helium, having an atmospheric concentration of about 7 parts in
Due to this paucity of $^3$He, the *in situ* decay of tritium produces significant increases in dissolved $^3$He: an increase detectable as an isotopic ratio anomaly. The measurement of the isotopic ratio of sea water helium rather than relying on absolute concentrations effectively cancels out natural fluctuations in helium concentrations due to air-sea interaction (bubble injection, radiative cooling, etc.) since any atmospheric

Figure 1. The Dreisigacker-Roether Source Function. Theoretical North Atlantic Surface Water tritium concentrations as a function of time (see text).
components added will have an atmospheric helium isotopic ratio. The "tritiogenic" helium is, of course, pure $^3$He.

In addition to the decay of tritium and the atmosphere, there are two other possible sources of helium. The first is radiogenic helium produced by decay of U and Th in sediments and the oceanic crust. Using a sedimentary concentration of 3 ppm for U, Th/U $\sim$ 4, and a characteristic sediment thickness of 0.5 km (Noltimier, 1974), one can compute that the contribution to the water column over a century "incubation" would be an order of magnitude smaller than current detection limits and three orders of magnitude smaller than the tritiogenic $^3$He "signal" seen in the thermocline. A similar calculation shows the oceanic crustal contribution to be even smaller. The second source in the Atlantic is the deep water injection of primordial $^3$He (see Jenkins et al., 1972). The only observed occurrence of this "juvenile helium" in the North Atlantic was near the Gibbs Fracture Zone at a depth of about 3200 m ($\sim$ 2.5°C, Jenkins et al., 1972; Jenkins and Clarke, 1976). Above this strata, no significant occurrences have been documented, although (see Sections 3 and 4, below) use of tritium and silicate budgets suggest a small (ca 2%e) primordial $^3$He contribution in the deep and intermediate waters. This clearly does not substantially affect the conclusions which may be drawn from the $^3$He distribution, but it can and must be accounted for.

The measurement of $^3$He enhances the information gained from tritium in two ways. First, for short time scales, the smallest detectable "isolation" time is decreased. This may be seen by noting that isolation of a water parcel from the surface mixed layer for two months will result in a 1% decrease in the tritium concentration, and a $^3$He increase (for a tritium concentration of 6 T.U.) of $1.4 \times 10^{-16}$ cc g$^{-1}$. The former change is comparable to current analytical accuracy ($\sim$ 1%); but it is below typical oceanic tritium "noise" (the change would be about 0.06 T.U., compared to a background variability of $\sim$ 0.5 T.U. in the Sargasso Sea). The latter change (in $^3$He) is, however, within current $^3$He analytical precision ($1 \times 10^{-16}$ cc/g). Thus $^3$He enhances minimum isolation sensitivity.

Second, for longer time scales, the $^3$He represents a coupled (to tritium by decay) tracer field with complementary boundary conditions: whereas tritium tends to be increased by surface contact, $^3$He is lost to the atmosphere. This adds a second dimension to the tritium picture and increases the sensitivity of models to exchange processes with the atmosphere. The source function for $^3$He is, of course, linked to tritium. The "chemistry" or boundary conditions are also known. It is inert, and all our data (see Section 5) clearly indicate that the "$^3$H-$^3$He clock" is zeroed in the mixed layer (no detectable $^3$He excess exists in the mixed layer).

3. Tritium-helium dating

The $^3$H-$^3$He dating technique, which was first exploited in the oceans by this investigator and others (Jenkins et al., 1972; Jenkins and Clarke, 1976), appears to
be applicable on the sub-decade timescale. The central concept to the technique is as follows: a surface water parcel, with a given tritium content and no $^3$He excess is advected below the surface and isolated from the atmosphere. $^3$He accumulates such that

$$\tau = \lambda^{-1} \log \left[ 1 + \frac{[^3\text{He}]}{[^3\text{H}]} \right]$$

(1)

For short time scales (a few years) this simplistic approach yields accurate results (cf. Jenkins, 1977). However, for time scales approaching that of the tritium tran-
sient, and when mixing occurs, the results break down due to mixing non-linearity (Jenkins and Clarke, 1976). Consider a simple box-mixing model, where an internally well-mixed box exchanges with the surface reservoir at a fixed rate, and assure the exchange takes place in the form of discrete, annual, instantaneous partial renewal events. At a time $t$, the $^3$H and $^3$He concentrations are given by

$$^3\text{H}(t) = \left(1 - \frac{1}{T}\right)^{^3\text{H}(t-1)} e^{-\lambda} + \frac{^3\text{H}_S(t)}{T}$$

$$^3\text{He}(t) = \left(1 - \frac{1}{T}\right) \left[ ^3\text{He}(t-1) + (1-e^{-\lambda}) ^3\text{H}(t-1) \right]$$

where $T$ is the characteristic renewal time, $\lambda$ is the decay constant (yr$^{-1}$) and $^3\text{H}_S(t)$ is the (time dependent) surface water tritium concentrations here taken as the Dreisigacker-Roether function (DRF). The model is initiated as a steady state system at 1950 assuming prebomb surface tritium levels of 0.2 T.U., and given by the continuous renewal balance:

$$^3\text{H}(O) = \frac{0.2e^\lambda}{(Te^\lambda-T+1)}$$

$$^3\text{He}(O) = (T-1) (1-e^{-\lambda}) ^3\text{H}(O),$$

and then "marched" through time. The $^3$H-$^3$He age is computed according to Equation (1). Figure 2 shows the response for such a box (flushing = 25% per year) to the tritium transient, and its computed $^3$H-$^3$He age ($\tau$). Figure 3 shows the $^3$H-$^3$He age response of several different boxes as a function of time. There are a number of notable aspects about this response to the tritium transient. The first is that the "initial" (i.e., steady-state) $^3$H-$^3$He age is not the same as the "flushing time", but is
lower. This depression is due to the mixing non-linearity mentioned above (surface water enters the box with a relatively higher tritium concentration, and is therefore weighted higher in the age averaging process, see Jenkins and Clarke, 1976). Quantitatively, we have

$$\tau = \lambda^{-1} \log [(T + (1-T)e^{-\lambda}]$$

for the steady state, where $T$ is the box flushing time and $\tau$ is the $^3$H-$^3$He age. Second, as would be expected, the transient depression is greatest at the time of maximum disparity between surface and interior tritium concentrations; that is around 1964 (see Fig. 1). Third, the recovery of the systems is more rapid for the faster flushing boxes. In fact, boxes with flushing times comparable to the tritium transient time scale (ca. 10 y) actually overshoot steady-state ages due to a “memory” of higher surface tritium concentrations.

Figure 4 shows the relationship between the measured $^3$H-$^3$He age ($\tau$) and the mixing rate for the box ($T$) at the present time (1978). For timescales of the order of 5 years or less, the $^3$H-$^3$He age is an accurate (~10% error) representation of the true age, but for longer timescales, the discrepancy becomes substantial. The encouraging aspect about Figure 4 is that, providing the box-mixing model holds, there is a one-to-one relationship between the measured $^3$H-$^3$He age and the “true” age. The result of the non-linearity at greater ages is then only a reduction of signal-to-noise. However, it is important to exercise caution in interpretation of $\tau$, especially with time scales longer than a few years.

4. Qualitative discussion of the data

A description of the sampling and measurement techniques is given in Jenkins et al. (1979), as well as a listing of the data used in this study. The helium isotope measurements were performed mass spectrometrically to an accuracy of 1.8‰ in $\delta$(He). Tritium measurements were performed by the $^3$He regrowth technique (cf. Clarke et al., 1976) to an accuracy of 1.5‰ with a detection limit of 0.06 T.U. Samples were taken from two cruises in 1974-1975 in the Sargasso Sea and from three reoccupations of the Panulirus Station (32.2N, 64.5W) during 1977.

The tritium and $^3$He results from the Panulirus stations are plotted as a function of potential density in Figures 5 and 6, respectively. Potential density rather than depth was used to enhance station to station correspondence.

The near-surface tritium data show a large scatter, with a general positive correlation between density and tritium (warmer water with lower tritium). This is likely a result of the Gulf Stream transporting into the Sargasso Sea warm, low-latitude, low-tritium water (the lower latitudes received less tritium). There is a maximum in tritium at $\sigma_o = 26.5$, which corresponds roughly to the Subtropical Mode water. This, as may be seen in the next section, may either be the result of a “memory” of previous, higher surface tritium values, or the vertical juxtaposition of northward
flowing low tritium water over southward flowing high tritium water. The tritium decreases dramatically through the upper main thermocline (27.4 > \(\sigma_0\) > 26.6) and reaches a minimum at about the depth of the oxygen minimum (\(\sigma_0\) ~ 27.4 to 27.5). This feature is not particularly surprising, since the \(O_2\)-minimum layer would be expected to be one of minimum exchange with the atmosphere (surface). This feature has been seen and reported on earlier (Ostlund et al., 1974). It should be noted, however, that the correspondence is not exact, and that the oxygen minimum lies slightly above the tritium minimum layer. This separation can be qualitatively explained by attributing an upward displacement of the oxygen minimum to the effect of a monotonically decreasing oxygen consumption rate with depth.

The deeper tritium maximum (\(\sigma_0\) ~ 27.7) has also been noted earlier (op. cit.), and is evidence of a lateral injection of tritium along this stratum. The source of this tritium remains problematic. Although the density of this maximum (\(\sigma_0\) ~ 27.7) is that usually associated with Mediterranean Water (cf. Sverdrup et al., 1942; Spencer,
1972) a simple calculation can show that its origin is more complex than this. The salinity anomaly here relative to the Wright-Worthington mean North Atlantic temperature-salinity curve (Wright and Worthington, 1970) is about .05‰ for this density surface. The corresponding tritium anomaly (relative to the "background" $^3$H-$\sigma_\theta$ curve) is about 1.2 T.U. The tritium content corresponding to the source salinity anomaly of 1‰ would be 24 T.U., clearly in excess of any past North Atlantic surface water concentration (see Fig. 1). Further, "Mediterranean Water" consists of only one third surface water (Zenk, 1975). Since both the true Mediterranean outflow and the entrained deeper water are low in tritium (paper in preparation), this requires even more unrealistic initial values. Consequently, a more direct, northerly source must be proposed for this deeper tritium maximum.

Below this depth, the tritium drops to "zero" at and below 2000 m. However, it is curious to note that the mean tritium level below 2000 m is 0.17 ± .09 T.U. (sample standard deviation), i.e., significantly above detection limit. The variations appear vertically coherent as well. This is not unexpected, since we have detected
Figure 6. $\delta(^{3}\text{He})$ vs. potential density anomaly for Panulirus Stations 1 to 3 (March-July, 1977).

substantial levels ($\sim 0.6$ T.U.) in the western boundary undercurrent (Jenkins and Rhines, 1980). Other workers (Roether and Münich, 1972) have detected significant levels of tritium in the deep waters in this area, and significant levels of an anthropogenic halocarbon (trichlorofluoromethane, or freon-11) have been detected (Hammer et al., 1978).

Qualitatively, the $^{3}\text{He}$ picture (Fig. 6) confirms this tritium pattern. Surface $\delta(^{3}\text{He})$ values cluster around isotopic equilibrium ($\delta(^{3}\text{He})_E \sim -12\%$) then increase to a maximum at about $\sigma_0 \sim 26.6$. This maximum is a result of a tradeoff between increasing isolation and decreasing tritium. The $^{3}\text{He}$ decreases to a minimum at $\sigma_0 = 27.4$ corresponding to the tritium minimum, then executes a deeper maximum which corresponds to the deeper tritium maximum. Beneath this, the $\delta(^{3}\text{He})$ values decrease to an average value (below 2 km) of $-6\%$, i.e., significantly above solubility equilibrium. Weiss (1970) determined a value of $-14\%$ for solubility equilibrium. More recently, Clarke (personal communication) has obtained a value of $-16.3 \pm 1.0\%$ for this temperature range, and Jenkins et al., 1979, using an ensemble of surface water $\delta^{3}\text{He}$ values, obtained $-17.4 \pm -0.28$ $T$ ($^\circ$C). Using a solubility
equilibrium value of $-16\%_{o}$, this gives an excess $^3$He content corresponding to the total in situ decay of 0.5 T.U., which is consistent with expected pre-nuclear era surface concentrations. Since this may have been potentially increased by a possible contribution of primordial $^3$He, and has been augmented by bomb-tritium decay, this may be regarded as an upper limit value. This estimate may be improved by accounting for the latter effect. By using a simple mixing model, this pre-bomb tritium level can be restricted to less than 0.4 T.U. A further refinement can be achieved by accounting for at least part of the primordial helium. Antarctic bottom water, as suggested by silica contents, contribute significantly to the hydrography below 200 m. Using data for this location (Metcalf, private communication) and assuming a "background" silicate content of 10 $\mu$M/Kg, we obtain a "silicate anomaly" of about 7 $\mu$M/Kg. Using a SiO$_2$-$\delta(^3$He) correspondence of 0.6$\%_{o}$ - Kg/$\mu$M (Jenkins and Clarke, 1976; Station 48, $T < 1^\circ$C), we obtain an AABW $^3$He correction 4$\%_{o}$. This reduces an upper limit estimate of pre-bomb surface tritium levels to 0.3 T.U.

Returning to the shallow $^3$He values, we can see that they show no consistent offset relative to expected solubility equilibrium. The average water ($\leq$10m) value is $-11.2 \pm 2.1\%_{o}$, and the average mixed layer value is $-11.7 \pm 1.2\%_{o}$. Whereas these are higher than the value predicted from Weiss' (1970) regression fit, his data for this temperature range (20$^\circ$C) averages to $-11.5\%_{o}$. Further, Clarke (private communication) has redetermined this value to be $-10.9 \pm 1.0\%_{o}$, and the Jenkins et al. (1979) analysis gives $-11.8\%_{o}$. However, it may be argued that the actual formation of a water mass may happen sufficiently rapidly that gas equilibrium may not be reached. For example, the work of Weiss et al. (1978) on Lake Constance and Broecker and Peng (1974) in the Bomex area suggest gas evasion rates of the order of a few meters per day. With expected deep mixed layer formation rates of a few days to a few weeks (cf. Medoc, 1970) this would clearly lead to disequilibrium. However, it should be noted that the gas exchange rate increases with wind speed (Kanwisher, 1963; Deacon, 1977), and with the high wind velocities associated with water mass formation (Medoc, 1970), high gas exchange rates would be expected. Further, the molecular diffusion coefficient of helium is relatively high (Boerboom and Kley, 1965), so that regardless of whether a stagnant or replacement film theory is used, evasion of helium should be substantially enhanced over other gases. Recent results in the Medoc area (Weiss and Jenkins, in preparation) indicate evasion rates for $^3$He of the order of 30 m/day during water mass formation, giving adequate exchange rates for gas equilibrium. Furthermore, the gas exchange results obtained may in fact be lower limit estimates due to possible lateral exchange with undegassed water. Consequently, gas exchange equilibrium remains a safe assumption.

Aside from the general trends, there is a significant temporal effect observed in the upper S.T.M.W. ($\sigma_{0} \sim 26.4$). The southward propagation of the winter's re-
newal event is visible not only in the dissolved oxygen (it increased by 0.4 ml/l from March to July at $\sigma_0 = 26.4$), but also in $^3$H and $^3$He: $\delta(^3$He) decreased by more than 15%, and the average tritium levels decreased by a fraction of a T.U. The reduction in $\delta(^3$He) corresponds to the ventilation and the decrease in tritium corresponds to the admixture of now lower tritium content surface water. The time lag between renewal (presumably February-March, cf. Warren, 1972; Worthington, 1959, 1978) and appearance of the $O_2$, $^3$H, $^3$He perturbations is indicative of net southward propagation rates of the order of 5 cm/sec. This homogenization is rapid relative to the ventilation time of this strata (ca 5y, see Jenkins, 1977 and Section 5).

5. A vertical model

There are a number of pathways by which tritium may enter the main thermocline. It is not immediately obvious that a one-dimensional data set (i.e. this monitoring station) may allow us to evaluate whether the tritium enters by cross-isopycnal or isopycnal exchange, or both. There is, in fact, a fundamental difference between the two modes: isopycnal exchange does not have the vertical coupling that local cross-isopycnal exchange should exhibit. The result is that the vertical distribution of tritium should evolve differently with time, depending on the exchange modes. Further, the added information obtained from $^3$He could be used. In a sense, the boundary conditions for $^3$He and $^3$H are complementary: those processes (that is exchange with the surface) which serve to increase tritium content, serve also to decrease $^3$He content. Thus a successful model of the exchange processes which brings the tritium into the main thermocline must take out the right amount of $^3$He.

In this section, a purely vertical model is discussed, and in the next section, a purely lateral model is presented. The former is discussed more for the sake of completeness than as a true interpretation. A purely vertical model cannot be extended deeper than the tritium minimum, since the presence of the deeper maximum demands lateral input. In principle, the shallow tritium maximum cannot be explained by strictly vertical processes either; but for the moment it is assumed that the observed structure is an artifact of an influx of low tritium surface water, and an average tritium concentration is used for the upper 400 m.

The equations of continuity are

\[
\frac{\partial}{\partial t} C(^3\text{H}) = K(z) \frac{\partial^2}{\partial z^2} C(^3\text{H}) + \left( \frac{\partial}{\partial z} K(z) - w(z) \right) \frac{\partial}{\partial z} C(^3\text{H}) - \lambda C(^3\text{H}) \tag{4}
\]

\[
\frac{\partial}{\partial t} C(^3\text{He}) = K(z) \frac{\partial}{\partial z} C(^3\text{He}) + \left( \frac{\partial}{\partial z} K(z) - w(z) \right) \frac{\partial}{\partial z} C(^3\text{He}) + \lambda C(^3\text{He}) \tag{5}
\]

Here, the advective flux divergence is assumed to be small. The pseudo-advec-
tive term \( \frac{\partial}{\partial z} K(z) \) results from a non-constant spatial dependence of the diffusivity. The sense of the expected variation in diffusivity (i.e., a decrease with depth) generates an effective downwelling velocity which may be significant relative to mean thermocline upwelling rates. For example, a decrease of 10 cm\(^2\)/sec over a depth range of 1 km would result in a downwelling pseudo-velocity of 30 m/y. This diffusive flux divergence term is particularly attractive in that if an exponentially decreasing \( K_e \) is used, coupled with a “typical” thermocline upwelling rate of 6 m/y, the sense of the “advective” flux (i.e., sum of the vertical advective flux and the divergence of diffusive flux) reverses smoothly at intermediate depths and provides a smooth transition from the apparent downwelling trend of the subtropical gyre (cf. Stommel and Webster, 1962) to the generalized upwelling required by thermocline models.

Noting that the above two equations are linear, the second may be rewritten as

\[
\frac{\partial}{\partial t} C(^3M) = K(z) \frac{\partial^2}{\partial z^2} C(^3M) + \left( \frac{\partial}{\partial z} K(z) - w(z) \right) \frac{\partial}{\partial z} C(^3M)
\]

(6)

where \(^3M\) is the mass three analog of “stable tritium” so that the \(^3\)He concentration may be obtained from

\[
C(^3\text{He}) = C(^3M) - C(^3\text{H})
\]

The general topology of solutions may be constrained by assuming that if the relation

\[
K(z) \propto \left( \frac{\partial \rho}{\partial z} \right)^{-1}
\]

approximately holds, then the conservation of density reduces to

\[
\frac{\partial}{\partial z} (wp) = 0
\]

which gives \( W(z) \) to be approximately constant, since \( \rho(z) \) varies by only a few percent over the range of interest. At this point, we will not attempt to satisfy surface boundary conditions with regard to the vertical velocity, since our major concern is to satisfy the observed property distributions in the main thermocline. As may be seen by the previous analysis and later results, the pseudo velocity term tends to dominate, especially near the surface.

A weighted-mean finite difference model (after Fiadero and Veronis, 1977) is employed, using a variable diffusivity and adapted for transient solutions. For enhanced stability, a variable metric was used, with a rough correspondence of

\[
\Delta h \sim \sqrt{K(z)}
\]

and an average resolution of 30 m, leading to a model accuracy of the order of \( 10^{-3} \)}
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The model was used to compute the steady-state $^3$H and $^3$He profiles for a pre-bomb surface value of 0.2 T.U. and $\delta(^3\text{He}) = -15\%$ (i.e., isotopic equilibrium with the atmosphere) for tritium and $^3$He, respectively, and for zero flux at the bottom boundary. The model was then forced by the Dreisigacker-Roether surface tritium function (1978, extrapolated past their 1975 limits).

What immediately became apparent was that no satisfactory fit could be achieved for a constant diffusivity. Qualitatively, the high diffusivity required for the penetration of tritium into the upper portion of the water column (the top 500 m) was too high to explain the tritium values at depth.

The next approach was to use a diffusivity exponentially decreasing with depth:

$$K = K_0 e^{-z/\lambda_0}.$$  \hspace{1cm} (7)

Since the depth scale of the thermocline is the order of 450 m, we fit the distribution of tritium to

$$K = K_0 e^{-z/450}$$

and find the best fit for $K \sim 10 \text{ cm}^2/\text{sec}$ and $\lambda = 5 \text{ m/y (upward)}$. However, even the best fit is not very satisfactory, for the same reasons as the constant $K$ case. If we lift the buoyancy flux constraint, a best fit to the 1977 tritium distribution is obtained with

$$K = 60 e^{-z/170}$$

The near-surface diffusivity is certainly artificial, but at the depth of the S.T.M.W. (~250 m), this gives a diffusivity of around $8 \text{ cm}^2/\text{sec}$ and within the main thermocline (~700 m) around $1 \text{ cm}^2/\text{sec}$. Clearly, other functionalities may be chosen, but the general result remains: a high diffusivity (ca $10 \text{ cm}^2/\text{sec}$) is needed in the upper water column and a lower value (ca $1 \text{ cm}^2/\text{sec}$) is required for the lower water column.

When this model is extrapolated to the past and compared with actual data obtained at this location (Fig. 7, solid line), the two begin to diverge, with the discrepancy becoming more pronounced for earlier times. The conflict does not appear to be severe, for it should be possible to tune the model to fit the profiles better, especially with regard to the lower boundary conditions. However, the historical data represents a weak constraint, since adequate reliable data is not available for the critical period during the onset of the tritium transient, when system response would have been the most telling.

Although the vertical model may adequately mimic the vertical distribution of tritium, it fails to simultaneously mimic the observed $^3$He distribution. This may be seen in a plot of $^3$He vs. tritium (Fig. 8), where the model predicts too little $^3$He in the upper waters and too much $^3$He in the tritium minimum. That is, in order to bring tritium down into the top of the thermocline by purely vertical exchange, the model loses too much $^3$He. Conversely, in order to prevent the tritium minimum
from "filling up", the model "holds in" too much $^3$He. The presence of the lower $^3$H, $^3$He maximum (not accounted for by this model) would aggrevate the latter conflict by diffusing even more $^3$He and $^8$H into the tritium minimum. For comparison, the two model curves discussed are shown. This failure of the vertical model exists regardless of the choice of the specific functionality of $K_v$.

The vertical model can be extended by assuming that lateral processes play a role above some critical depth, but that vertical processes predominate below this depth. The critical issues that must be decided are the choice of this depth and the nature of the boundary conditions applied to this "quasi-vertical" model. The former decision is relatively straightforward: the upper boundary depth should be in the vicinity of the tritium maximum (390 m, $\sigma_o = 26.5$), which coincides with the depth of the subtropical mode water. This choice is attractive in that it coincides to the minimum surface water temperature generally observed south of the Gulf Stream in late winter.

The exact boundary conditions applied to the model require more careful consideration. An initial approach would be to fix the concentrations of the boundaries according to some time-varying laterally-controlled mixing response to the surface forcing (specifically according to Equations (2) and (3), see Section 6). However, this approach would be unrealistic in that it does not entail a continuity of (tritium or $^3$He) flux. That is, it does not conserve mass. The correct coupling of the vertical and lateral processes is embodied in the following equations:

$$\frac{\partial}{\partial t} C(\text{H}) = K \frac{\partial^2}{\partial z^2} C(\text{H}) - w \frac{\partial}{\partial z} C(\text{H}) - \lambda C(\text{H}) + S(z) (C_{8}(\text{H}) - C(\text{H}))$$

(8)

$$\frac{\partial}{\partial t} C(\text{He}) = K \frac{\partial^2}{\partial z^2} C(\text{He}) - w \frac{\partial}{\partial z} C(\text{He}) + \lambda C(\text{H}) - S(z) C(\text{He})$$

(9)
where \( S(z) \) is a depth-dependent surface exchange coefficient and \( C_S (^3\text{H}) \) is the time-varying surface tritium concentration. The surface exchange coefficient can be adjusted to give the appropriate \(^3\text{H} \) and \(^3\text{He} \) concentrations at and above the upper boundary at 390 m, and be made zero below that depth. Further, this allows the use of the deeper tritium/\(^3\text{He} \) maximum at 1000-1200 m as the lower boundary, permitting a more realistic representation of the tritium/\(^3\text{He} \) minimum in between and presumably its evolution with time.

Since the density gradient in this more restricted depth range is relatively constant, use of a variable diffusivity would not be justified, so that a constant diffusivity is used. Whereas the diffusive flux divergence term in Equations (4) and (5) dominated the advection term, this is not the case here, so that some care must be taken in parameterizing the vertical velocity. Examination of the temperature profiles shows a negative curvature (i.e., temperature gradient increasing with depth) at about 400 m and a positive curvature (temperature gradient decreasing with depth) around 1000 m and below. If vertical processes prevail, this suggests a reversal of the vertical velocity component from downward above the thermocline (driven by Eckman suction) to upward below the thermocline (global thermohaline upwelling). The zero velocity point would be at about 700 m where the temperature profile is linear. Over the limited range between 700 m and two boundaries, it is possible to estimate the vertical velocity according to:
\[ \Delta w = \frac{\beta}{f} \int_{\Delta z} v dz, \]

where \( v \) is the meridional component of the horizontal velocity. Using Worthington's transport charts (Worthington, 1976, Figs. 24, 26, 29), an average southward velocity of the order 1 cm/sec is obtained over the depth range 400-1200 m. This yields an estimated vertical velocity of

\[ w = 1.62 \times 10^{-4} - 2.32 \times 10^{-7} z \]

where \( w \) is in cm/s (again positive downward) and \( z \) is in meters. This could be expected to be relatively accurate over the depth range prescribed, but the surface extrapolated value of \( 1.62 \times 10^{-4} \) cm/s (downward) is encouragingly close to Leetmaa and Bunker's (1978) estimate of the annual mean vertical velocity at the base of the Ekman layer (cf. their Fig. 2). Thus the downward vertical velocity of \( 7 \times 10^{-5} \) cm/sec (about 20 m/y) at the top of the vertical domain (400 m) is obtained, whereas a substantially larger upward vertical velocity of \( 1.2 \times 10^{-4} \) cm/sec (about 35 m/y) is obtained at the base of the domain (1200 m). This latter value is disturbingly larger than expected from considerations of general thermocline balance, but it must be recalled that a general trend of decreasing \( v \) with depth would be expected so that this number is likely a substantial overestimate. Clearly, however, the point is academic since the fluxes are indeed small due to low concentrations, and as the analysis proceeds, it will become clear that the critical region of judgment will be at and above 700 m, where the vertical velocity is either small or downward. Attenuation of the vertical velocity below this depth does not substantially affect the results.

Figure 9 compares the data to model calculations, the latter obtained by adjustment of the vertical diffusivity and surface exchange coefficients to achieve the best fit to the tritium profile! For comparison, the vertical velocity was scaled by \( \pm 50\% \) and the results included.

What is again evident is the clear and irreconcilable differences between data and model. It was not possible to achieve a very satisfactory fit to the vertical tritium distribution, because the 400 m maximum in tritium is eroded by advective and diffusive loss to the thermocline waters below. If the advection-diffusion "strength" is decreased so as to preserve the feature, inadequate tritium penetration into the thermocline results. If the surface exchange coefficient is increased (to compensate for the added flux required), the maximum would wash out due to upward exchange. However, when even the best possible fit is achieved, the resultant \(^8\)He profile is still inadequate. The discrepancy is qualitatively explained by the same arguments as those applied to Figure 8, and can be readily seen by the simple schematic shown in Figure 10.

Tritium that is brought into the "box" (i.e., the STMW at and above 400 m) by both lateral and vertical processes supplies the underlying thermocline with tritium
Jenkins: Tritium and $^3$He in the Sargasso Sea

Figure 9. Observed and quasivertical model $^3$H vs. $^3$He for the upper 1 km. Symbols represent the data, solid line is the best fit for $K_v = 1 \text{ cm}^2/\text{sec}$, and $w$ as in the text. Broken line is for $\frac{1}{2} w$ and dotted line is for $1.5 w$.

(see Fig. 10a). Thus the rate of supply to the “box” must be adequate to build up and maintain the STMW $^3$H inventory against advective and diffusive loss to the water column below. This “balance” is very poorly approached by the quasi-vertical model. Even so, if one contends that the general “flavor” of the $^3$H penetration is achieved, the $^3$He budget is not satisfied. The $^3$He that is generated within this “box” by tritium decay is lost by all three pathways (see Fig. 10b), so that the sum total of the loss mechanisms must not be so large as to deplete the STMW $^3$He inventory below the observed levels. Although to some extent, the surface exchange parameter can be adjusted to at least obtain the approximate $^3$H inventory, as we have done in the model calculations, this serves to deplete to $^3$He inventory. Clearly, the quasi-vertical model violates these inventories. A detailed analysis of the model’s response to variations of the parameters (advection, diffusion and lateral exchange) shows that the “balance” cannot be improved.

In fact, the vertical advective-diffusive transport mechanisms serve to separate the $^3$H and $^3$He maxima spatially, and the $^3$He maximum is attenuated, broadened, and “washed” down into the water column. This is, in part, an advective effect, but also is driven by the substantially greater $^3$He gradient above the maximum compared to the more gentle gradient below. The spatial proximity and magnitude of the $^3$H and $^3$He maxima actually observed at about 400 m allows a crude limiting calculation on the magnitude of vertical mixing in operation. Assuming “stationarity” of the $^3$He maximum allows constraints to be made on the relative fluxes. Since the
Figure 10. A schematic representation of (a) the tritium budget and (b) the $^3$He budget for the subtropical mode water at 300-400 m depth. Note that whereas tritium enters via two routes and leaves by only one, $^3$He leaves by all three routes. A successful budgeting must carry tritium into the STMW and the thermocline, yet at the same time not remove too much $^3$He.

$^3$He maximum has grown from virtual non-existence and has not changed in magnitude by very much over the past few years (see Jenkins et al., 1979), this stationarity assumption is not seriously violated. We have then
\[ \frac{\Delta C}{\Delta t} o = K \left[ \frac{\Delta C}{\Delta z} \right] + - K \left[ \frac{\Delta C}{\Delta z} \right] + U_+ (C)_+ - U_- (C)_- + \lambda(\text{H}) - S(C)_o \]

where + and − are evaluated at the top and bottom of the “box” respectively. The vertical extent of the box would be about 150 m, as determined by the width of the \(^{3}\text{He}\) maximum and the effective value of the surface exchange coefficient can be no less than .09 y\(^{-1}\) (lesser values would yield lower \(^{3}\text{He}\) inventories due to lower \(^{3}\text{H}\) inventories, see Fig. 11), and we use a mean \(^{3}\text{H}\) inventory of 5.1 T.U. from the “box”. We then obtain

\[ \frac{\Delta C}{\Delta t} = o = -K \times 1.7 \times 10^{-8} - 2.0 \times 10^{-9} + 9.2 \times 10^{-9} - 6.6 \times 10^{-9} \]

which yields

\[ K \leq .05 \text{ cm}^2/\text{sec} \]

A careful consideration of the relative uncertainties in each of the terms in the above equation suggests that a firm upper limit to \(K\) is no more than 0.1 cm\(^2\)/sec, but in many respects, the terms are conservatively estimated, so that the true vertical diffusion coefficient may indeed be substantially less than this upper limit.

A further argument against vertical models is that the high values of the vertical diffusivity (1-10 cm\(^2\)/sec) needed to bring the tritium down into the thermocline by purely vertical processes are substantially higher than what might be expected from other observations as well. Armi (1978) has observed deep, bottom detached mixed layers in the Sargasso Sea (depth \(\sim 4\) km) the presence of which are consistent with vertical diffusivities of the order of .01 cm\(^2\)/sec. Gregg (1977) reports Cox numbers of the order of 10 for the main thermocline of the North Pacific subtropical gyre (cf. Fig. 8 of op cit.), corresponding again to \(K_v \sim .01\). Using similar measurements, Gargett (1976) reached a similar conclusion, with a model value of \(K_v\) around .02 to
Also, using temperature and salinity fine-structure measurements, Schmitt and Evans (1978) estimated a vertical diffusivity of the order of 0.1 cm$^2$/sec due to salt fingering. Thus observations of oceanic thermohaline structures suggest substantially lower vertical diffusivities than required by this model.

A further example which suggests that vertical exchange is not a dominant process in the thermocline may be seen in the distribution of salinity anomaly in the Mediterranean water in the North Atlantic. Needler and Heath (1975) noted that in all but one instance, the vertical scale (i.e., thickness) of the Mediterranean salt tongue decreased away from the Straits of Gibraltar. They suggested that horizontal rather than vertical processes were responsible for transport of salt in the Mediterranean water. Pingree (1972) and Hayes (1975) came to similar conclusions based on deepwater small-scale $T,S$ structures associated with Mediterranean water in the Atlantic. The small and medium scale structure arguments therefore suggest a predominance of lateral over local vertical exchange. In some respects, this may be regarded as a lower limit estimate of the strength of lateral processes in affecting large scale ocean structure in that the scale effect on lateral diffusion (see Stommel, 1949, Okubo, 1971) will enhance lateral exchange on the larger scales.

Based on near bottom vertical profiles of $^{222}\text{Rn}$, Sarmiento et al. (1976) obtained typically values of the order of 10 cm$^2$/sec for $K_v$ for the deep water. Assuming constant buoyancy flux, they extrapolated their results to the main thermocline to obtain a $K_v$ of .01 to .05 cm$^2$/sec. Aside from the question of the constant buoyancy flux assumption, their extrapolation may be regarded as an upper limit to the true vertical diffusivity since they were, in effect, in their models projecting both lateral and vertical exchange onto the vertical dimension. Indeed, they were unable to model a substantial fraction of their data set due to clearly laterally induced structures in the $^{222}\text{Rn}$ profiles. Rooth and Ostlund (1972), based on the apparent stationarity of the deep tritium distribution in the Sargasso Sea main thermocline over their measurement period (1968-1972), computed a net diffusivity (again projected on the vertical axis) of less than 0.2 cm$^2$/sec. Since the tritium may have entered the thermocline laterally, the Rooth and Ostlund estimate represents at best an upper limit to the cross-isopycnal exchange coefficient. Thus the tracer data available ($^{222}\text{Rn}$ and $^3\text{H}$) again suggest substantially lower vertical diffusivities than required by this model.

Schuert (1970) injected Rhodamine Dye at a depth of 300 m in the North Pacific, and observed no detectable vertical mixing over a period of 72 hours. The dye patch was 2 m thick, and assuming a detection limit for the scale height change of the order of 50% (1 m), an upper limit of .04 cm$^2$/sec is obtained. Although this clearly is not a statistically substantial conclusion, since vertical mixing is likely to be heterogeneous both temporally and spatially, this result is in agreement with the arguments made here.

In addition, it is not entirely clear that a vertical model is consistent with even
steady-state tracer distributions. As pointed out by proponents of one-dimensional advective-diffusive models (Munk, 1966; Craig, 1969), and as may be demonstrated with simple algebraic manipulation of the appropriate equations, stable-conservative tracers must be linear function of one another in the "advective-diffusive subrange". For temperature and salinity, this appears to hold over a limited range (8° to 16°C), but for silicate-temperature, there is a pronounced curvature. While it may be argued that silicate is not generally conservative, there is no abundant source of siliceous tests or skeletal material in this area (cf. Sverdrup et al., 1942, Broecker, 1974), so that significant in situ production of silicate at intermediate depths is unlikely. This is further corroborated by the observed 3He-SiO₂ relationship in the intermediate waters of the South Atlantic (Jenkins and Clarke, 1976). In summary, a constant diffusivity vertical model cannot satisfactorily generate the observed vertical tritium distribution in the Sargasso Sea. When a depth variable diffusivity is used, the present (1977) tritium profile can be generated, but the resultant model 3He distribution is clearly at variance with observations, as are past tritium profiles. Incorporation of lateral exchange to a depth of 400 m does not improve the agreement. The vertical diagnostic models used for tritium are also characterized by vertical diffusivities of the order of 1-10 cm²/sec, whereas most quantitative studies, including a detailed 3He flux balance calculation performed here, suggest vertical diffusivities of the order of 0.1 cm²/sec or less. Consequently, it can be concluded that vertical exchange processes are not primarily responsible for ventilation of the main thermocline in the Sargasso Sea.

6. A lateral transport model

If cross-isopycnal exchange can be ruled out as primarily responsible for the penetration of tritium into the main thermocline, then isopycnal or lateral processes must dominate. Further, the existence of maxima in the vertical tritium distributions cannot have been produced by vertical processes alone. Armi (1978) proposed a scheme whereby effective cross-isopycnal exchange was achieved by topographically generated mixed layers at the boundary which were then laterally transported into the interior along isopycnals. One possible approach which suggests itself is to consider an analogous process to apply for the main thermocline. In this case, the "boundaries" are late winter mixed layers, through which ventilation is achieved by relatively deep convective or turbulent mixing. When the seasonal pycnocline is formed in early April, this "winter water" is isolated from the atmosphere. Transport then occurs along isopycnals into the interior. In principle, this is the notion of Wust (1936), Montgomery (1938) and Iselin (1936) that the main thermocline is not formed by a balance of vertical mixing and upwelling, but by isopycnal processes connecting with the surface. The resultant "ventilation time" (in years) of a given density horizon (σ) will be
\[ T(\sigma) = \frac{V_I(\sigma)}{V_{WML}(\sigma)} \]

where \( V_{WML}(\sigma) \) is the volume of winter mixed layer produced annually of that particular density type, and \( V_I(\sigma) \) is the volumetric census of that density type in the interior. Implicit within this approach is that the ventilation period (i.e. that time during which the winter mixed layer is ventilating to the atmosphere) is small relative to 1 year, and that the interior is well-mixed on the timescale of the ventilation time \( T \). The former is easily justified since most of the winter is spent eroding the seasonal thermocline, so that the deep mixing event occurs over a period of a few weeks to a month (cf. Warren, 1972; Medoc, 1970). If the transport into the interior of the late winter mixed layer (LWML) during the ventilation period were to be taken into account, the effective ventilation rate predicted by the above equation would be enhanced. However, the present approach will be to parameterize this enhancement into a somewhat larger \( V_{WML} \). A simple estimate of the magnitude of this effect based on time spans is of the order of 10%.

The validity of the latter assumption is dependent essentially on the recirculation rate of the Sargasso Sea relative to the ventilation timescale. Using the geostrophic transport charts constructed by Worthington (1976, Figs. 24, 26, 29, 42), we find characteristic recirculation times of the order of 1 to 5 years. Similarly, Wunsch (1978) applied linear inverse theory to the general geostrophic circulation of the North Atlantic and obtained recirculation timescales varying from a few years (his model 1) to less than a decade (his model 2). Providing that ventilation time is in excess of about a decade, the homogeneity assumption is valid. This can be further tested by examination of the latitudinal distribution of tritium between 30N and 36N at about 60W for 1974 (Fig. 12). (Unfortunately, the \( ^3 \)H data was obtained only for \( \sigma_T \leq 27.6 \). The error bars given for each density surface correspond to the uncertainty associated with random sampling within a given density interval, the uncertainties being largest for the strongest tritium-pycnal gradient. Although there are indications of significant differences across the Gulf stream (G.S.), i.e., between slope water and the interior, there are no significant trends with latitude within the Sargasso Sea itself. This again confirms the earlier support based on geostrophic computations, i.e., the interior appears homogeneous on timescales comparable to the time elapsed time the surface water tritium transient, viz. ca. 5-10 years.

At this point, there are two possible mechanisms of communication between the LWML and the interior. If one assumes an exchange process, e.g., mixing or eddy-diffusion, then the requirement of mass conservation is automatically satisfied, since no mass transport occurs. However, this may be physically unrealistic since mass transport must occur due to Ekman pumping. Stommel (1979) has proposed a scheme whereby the Ekman regime is interfaced to the geostrophic flow below,
Figure 12. Average tritium concentrations vs. latitude. Samples taken from two cruises in November, 1974 and March, 1975 along 67.5 and 58.5W, respectively. Error bars are uncertainties associated with random sampling within a density internal when tritium is density dependent. G.S. = Gulf Stream position determined by XBT data.

which in many respects satisfies the upper boundary conditions of this model. In order to conserve mass, this model must be essentially open-ended, where the geostrophic stream lines leave the subtropical gyre. Reality may lie somewhere between these two cases, but within the framework of the model being discussed, they are mathematically the same, and no attempt at discerning between the two can be made.

Basically, the model resembles a box model extended to the continuum. Dealing with absolute fluxes and volumes can be avoided by a normalization as in Equation (8). This avoids the question of defining exactly the volume of the interior reservoir, as governed by both the spatial domain and the potential density interval. The governing equations are then (2) and (3), forced by the DRF, and the variable to be solved for is the ventilation timescale (inverse ventilation rate) \( T \). This approach also avoids the detailed parameterization of the processes responsible for lateral transport away from the mixed layers: whether the actual process is advection (Ekman pumping), Fickian turbulent diffusion, or some non-Fickian transport, the impedance limiting step is the amount of LWML produced annually. The formation of interior water types within the main thermocline then becomes a statistical process of which the temporal stability, that is the insensitivity of characteristics to variations in meteorological forcing (cf. Warren, 1972), follows quite naturally from the ventilation time \( T \) and whether it is long relative to the timescale of short term climate fluctuations.

Operationally, this can be accomplished by constructing a nomograph of \(^3\)H, \(^{3}\)He and \( \tau \) (\(^{3}\)H-\(^{3}\)He age) as a function of \( T \) at the time of sampling (July, 1977). This nomograph (Fig. 11) demonstrates the contribution of \(^{3}\)He as a tracer: for timescales
Table 1. Lateral transport model calculations.

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<th>Depth (m)</th>
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<th>$\Delta^{3}\text{He}$ (cc/g)</th>
<th>$\tau^{3}\text{H}$ (y)</th>
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<td>68</td>
<td>10</td>
<td>73</td>
<td>71</td>
<td>1.28</td>
</tr>
<tr>
<td>27.81</td>
<td>1600</td>
<td>1.00</td>
<td>1.81</td>
<td>96</td>
<td>20</td>
<td>101</td>
<td>99</td>
<td>1.21</td>
</tr>
<tr>
<td>27.82</td>
<td>1700</td>
<td>0.70</td>
<td>1.38</td>
<td>140</td>
<td>30</td>
<td>156</td>
<td>150</td>
<td>1.15</td>
</tr>
<tr>
<td>27.83</td>
<td>1800</td>
<td>0.50</td>
<td>0.95</td>
<td>200</td>
<td>50</td>
<td>310</td>
<td>260</td>
<td>1.10</td>
</tr>
</tbody>
</table>

less than about a decade, tritium becomes relatively insensitive to the ventilation rate. $^{8}\text{He}$, however, shows an enhanced sensitivity in the subdecade range. Although the quantity $\tau$ (the $^{3}\text{H}$-$^{8}\text{He}$ age) does not add new information in a canonical sense; for the short timescales it allows us to "scale" the $^{8}\text{He}$ to the actual tritium concentration rather than model tritium values.

To characterize the observed tracer fields, an average $^{3}\text{H}-\sigma_0$ and $^{8}\text{He}-\sigma_0$ curve was constructed, and the predicted ventilation time was obtained from the model calculations and summarized in Table 1. The uncertainty in the time computed from this model is obtained by model response to variation of the mean $^{3}\text{H}$, $^{8}\text{He}$ curves within estimated bounds of uncertainties, and model response to variation of the forcing function within constraints of existing reliable data.

The resultant ventilation age curves projected on a depth scale are shown in Figure 13. The reliability of the $^{8}\text{He}$ age computations is poor for short timescales due not only to data uncertainties, but also to model forcing function uncertainties. Further, the susceptibility to short term meteorological variability is greater for the
faster responding systems. Consequently, the more reliable indicator of at least "instantaneous" ventilation rate would be the \(^3\text{H}-^3\text{He}\) age, allowing for (at the longer timescales) the non-linear response, as in Figures 4 and 11. For densities greater than 26.7 (i.e. apparent timescales greater than about a decade), the \(^3\text{H}\) and \(^3\text{He}\) age values are more reliable.

What is immediately apparent is that the original assumption of internal homogeneity is vindicated, since the timescales estimated far exceed the recirculation timescale for the Sargasso Sea gyre. The rapid response in the shallow waters (\(\sigma_\theta \sim 26.4\) to 26.5) to the winter renewal event, as noted earlier, suggests that the assumption may hold for shallower depths as well. The most encouraging aspect of Figure 13, however, is the strong mutual agreement between \(^3\text{H}\) and \(^3\text{He}\): within the uncertainties of the calculations, the two tracers yield the same results. There does, however, appear to be a small systematic difference between the two tracers around \(\sigma_\theta = 27.5\), with \(^3\text{He}\) predicting shorter ventilation times. This can be explained by a small admixture of Antarctic water: a tongue of high silica, high \(\delta(^3\text{He})\)
water can be seen extending northward from the Antarctic (Jenkins and Clarke, 1976). Since there exists a linear relationship between silica and excess $^3$He (op. cit.) it should be possible to correct for the Antarctic high $^3$He influence using the predetermined $^3$He-SiO$_2$ relation. Although silicate data are not available for these stations, comparable station data (Metcalf, private communication) indicate a silicate excess at this density of a few micromoles per kilogram. It is not possible to determine the absolute excess since “background” silicate values cannot be defined; but this may give an estimate of the magnitude of the effect. A silica anomaly of 3 $\mu$M/kg corresponds to an Antarctic component of $0.28 \times 10^{-15}$ cc/g $^3$He. If the $^3$He is reduced by this amount, the $^3$He age is increased (see Fig. 11) by 12 y, bringing it into even closer agreement with the $^3$H age. In this sense the uncorrected $^3$He age may be regarded as a lower limit to the ventilation age, although the error is of the order of the model uncertainty. The best approach is thus to take the average of the $^3$H and $^3$He ages as the best estimate for the ventilation time, except near the surface, where the $^3$H-$^3$He age represents a more accurate estimate.

Further the correspondence between the predicted profiles and the actual tritium data is very good for the past (Fig. 7, dashed line), and agreement which was achieved without parameter optimization with regard to those data sets.

Table 2 compares the lateral transport model results ($LTM \tau$) with results obtained by Worthington (1976, Table 9) for a simple box flushing model for the North American Basin. As might be expected, the $LTM$ ages are systematically lower than the box estimates. This is due to the fact that the box estimates represent the purely advective component of the ventilation process. In a way, the ratio of the two ages represents a Peclet number in a non gyre-resolving sense, but this interpretation is complicated by the fact that the degree of local surface expression of the isopycnals must be taken into account. This is especially clear for the shallowest stratum ($>17^\circ$), but there does appear to be a tendency toward a more advective renewal in deeper strata. This more advective nature is to some extent supported by the slight latitudinal tritium gradient which appears at this density surface (see Fig. 12). This does not relax our original assumption of internal homogeneity, since the gradient is not substantial. The only major and inexplicable divergence between these estimates is for the 7-4$^\circ$C range. This difference is largely driven by the appearance of the mid-depth tritium maximum, while Worthington does not identify any significant influx or efflux for this “reservoir”. In order to account for the observed ventilation by purely advective processes, one would require a transport of about 2.5 Sv, which is difficult to reconcile with the Worthington scheme. The question may arise as to whether, assuming Worthington’s budgets are accurate, this observed flux may be diffusive. Characteristic (estimated) values of $K_H$ for this depth/density range are of the order of $1-3 \times 10^7$ cm$^2$/sec (e.g., see Needler and Heath, 1975). Assuming a characteristic distance of the order of 2000 km (roughly the distance to 55N), we obtain a characteristic time of the order of 65 years; i.e.,
Table 2. A comparison of LTM results with Worthington's (1976) residence times.

<table>
<thead>
<tr>
<th>Temperature range</th>
<th>(C°) Density</th>
<th>LTM ( \bar{\tau} )</th>
<th>Worthington ( \tau_w )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>range (%)</td>
<td>(y)</td>
<td>(y) ( \tau_w/\bar{\tau} )</td>
</tr>
<tr>
<td>&gt;17</td>
<td>&lt;26.5</td>
<td>~1</td>
<td>34</td>
</tr>
<tr>
<td>17-12</td>
<td>26.5-27.0</td>
<td>16</td>
<td>38</td>
</tr>
<tr>
<td>12-7</td>
<td>27.0-27.5</td>
<td>54</td>
<td>76</td>
</tr>
<tr>
<td>7-4</td>
<td>27.5-27.82</td>
<td>74</td>
<td>&gt;387</td>
</tr>
</tbody>
</table>

comparable to the LTM estimates for this stratum. Perhaps a more realistic assessment of the source of this tritium maximum can be obtained by considering the general anticyclonic circulation at this depth pictured by Worthington. A degree of "cross frontal exchange", executed by mesoscale processes under the high eddy energy region of the Gulf Stream extension (cf. Dantzler, 1977) could tag this water which would then be advected anticyclonically southward along the western flank of the Mid Atlantic Ridge and then southwestward. The gradients in tritium expected from this scenario would be increasing eastward with a sudden decrease across the M. A. R. since the geostrophic flow at this stratum (following surfaces of constant \( f/h \)) would be deflected strongly southward at the Ridge. This is borne out both in the NAGS tritium results of 1972, (Ostlund et al., 1977) and by more recent Western Boundary current results (Jenkins and Rhines, 1980). Consequently, one could then reconcile this to Worthington's budgets if some degree of lateral diffusion were involved.

The apparent success of the LTM suggests that isopycnal rather than local vertical transport processes are mainly responsible for the transport of material into the main thermocline and deeper. It is possible to assess on the basis of the present tracer data an upper limit to magnitude of vertical diffusion, and with the other evidence discussed earlier (Section 4) that true vertical diffusivities are of the order of 0.1 cm²/sec or less. Since the diagnostic vertical models require diffusivities of the order of 1-10 cm²/sec, we can conclude that actual vertical transport contributes from 1-10% of the mass balance. Certainly a more realistic model than the LTM would include vertical exchange as a perturbation, but such a refinement appears unwarranted on the face of the data at hand. More importantly, an attempt should be made to understand and correctly characterize the nature of the impedance limiting process; i.e., the winter mixed layer ventilation and in-mixing.

No contention should be made as to the general applicability of the LTM, particularly to regions of less rigorous recirculation; e.g., the eastern subtropical Atlantic. The choice of the Sargasso Sea as a "test area" is obviated by its relatively low impedance to lateral transport, so that the impedance limiting processes are accentuated. Extension of the model to higher impedance regions must include explicitly lateral transport by advection or diffusion.
Further, the low $K_v$ has important implications with regard to that class of thermo-cline models which require large (>0.1) vertical diffusivities, e.g., the Blandford (1965) model, and suggests that the non-diffusive models, e.g. Welander's (1959) or Hodnett's (1978) models, or models incorporating isopycnal diffusion (see Veronis, 1969), may be more realistic. Truly isopycnal transport may carry heat into the interior (with an associated salt flux), but by definition cannot transport buoyancy to balance a uniform upwelling. Any degree of non-isopycnal horizontal mixing will also destroy buoyancy due to the negative curvature of density in $T$-$S$. Consequently, the results obtained here preclude a uniform upwelling, at least in the subtropical gyre. In this respect, the low value of $K_v$ obtained here does not conflict with the high value of $K_v$ (~1 cm$^2$/sec) obtained from the vertical thermocline balance equations, but instead indicates that the bulk transport is localized rather than diffuse (cf. Needler, 1979), and is more in line with the bulk of estimates based on fine-scale thermohaline structure and large-scale tracer studies (see Garrett, 1979, for a recent summary).

7. Extension of the lateral transport model: apparent oxygen utilization rates

Within the confines of the lateral mixing model, it is possible to compute the rate of oxygen consumption. It is perhaps more realistic to refer to this consumption rate as an “apparent utilization rate” since the observed oxygen concentrations are the result of not only in situ consumption, but also, (in shallow waters at least), photosynthetic production. For our purposes, the latter is not important in deeper waters, except that it shifts the “starting point” in the AOU calculation. Since surface and near surface waters are generally within 5 to 10% of saturation, this remains a small perturbation.

The AOU was characterized by a smoothed AOU vs. $\sigma_\theta$ plot for the stations, and summarized in Table 1. The AOU rate (AOUR) was then computed using the best estimate ventilation rate, and is plotted vs. depth in Figure 14. For comparison, Riley's (1951) estimates (open symbols) have been included. The error bars correspond the uncertainties associated with both model ages and AOU values, while for the Riley data, the uncertainties are obtained from his Table 10.

The lateral mixing model AOUR values decrease smoothly from near surface rates of the order of 0.5 ml/l/y to a rate of $4 \times 10^{-3}$ ml/l/y at 1800 m, with an e-folding depth of about 350 m. The scale height of this AOUR decrease is remarkably similar to Wyrtki’s (1962) scale height for oxygen consumption rate, based on a two-dimensional model.

There appears to be a smooth structuring of the mean trend, with a tendency toward an inflection to lower values at about 1000 m. This cannot be explained by the effect of the $^8$He age depression due to the Antarctic influence discussed in the previous section. Within the uncertainties of the model assumptions, it is better not
Figure 14. AOUR vs. depth for the lateral mixing model. Solid symbols represent model results and triangles are Riley's (1951) calculations. Open circle is an estimate based on $^3$H-$^3$He dating (Jenkins, 1977).

to overinterpret this. What is evident, however, is that the oxygen minimum is more a result of physical processes (i.e., poor ventilation) rather than being a layer of maximum consumption rate. The expression of the minimum is in part due to the presence of underlying "more ventilated" water and in part due to the tradeoff between increasing isolation and decreasing consumption rates.

The agreement with Riley's (1951) and Jenkins' (1977) estimates is qualitatively good above the oxygen minimum, but Riley's estimates diverge to lower AOUR values below this depth. These differences may be rationalized in a number of ways. Riley's values were based on steady-state, three-dimensional advective diffusive modelling, using geostrophic velocities, the principle weakness of which lies in the "depth-of-no-motion" assumption. The errors introduced by this assumption will be smallest near the surface and greatest in deeper strata. Also, the Gulf Stream transport he uses (41 Sv at 27N, which is essentially the total transport since his recirculation occurs south of 20N, see Figs. 17-21, *op cit.*) is a factor of two or three lower than more recent estimates (Worthington, 1976, 1978; Schmitz, 1977). Further, his values of lateral diffusivity ($0.2-3.0 \times 10^6 \text{ cm}^2/\text{sec}$) are at least a factor of two or three lower than that used in current non-eddy resolving models ($0.6-3.0 \times 10^7 \text{ cm}^2/\text{sec}$, Kuo and Veronis, 1970, 1973; Needler and Heath, 1975). In this respect, it could be said that Riley's abyssal ocean may be regarded as underventilated, requiring correspondingly lower AOUR to explain the observed AOU values.

Riley's estimate of the deep water (2500 to 4000 m) AOUR is about $10^{-4}$ ml/l/y, although he performed an upper limit calculation based on the mean AOU (2 ml/l), a bottom water production rate (6 Sv) and the NADW volume census, and obtained a value of $2 \times 10^{-3}$ ml/l/y. It is this upper limit value which has proven acceptable for world ocean modellers (Kuo and Veronis, 1973). A crude extrapolation of this present work gives a value of .0005 at 2500 m which compares moderately well with a similar calculation for the North American basin which can be performed using Worthington's estimated residence time (130 y for $T < 4^\circ$) and an average AOU
of 1.2 ml/l. However, it is not entirely clear that the two results (this work and the Worthington type calculation) can be realistically compared, not only on conceptual grounds, but also on the validity of the extrapolation of this work to deeper strata.

The extrapolation may not have been as unrealistic as it might appear. First, it should be realized that strictly vertical processes, viz., hydrostatic pressure and particulate flux (grazing), do play a role in controlling respiration rates. Second, the actual persistence of the trend over 1800 m of depth range is suggestive of some degree of vertical control. The possibility of the vertical control begs the question of the applicability of the LTM to compute the AOUR. In defense of this, it should be noted that formation of the LWML takes place very quickly (Medoc, 1970) and is associated with high gas exchange rates (Weiss and Jenkins, in preparation). Coupled with the intense recirculation within the Sargasso Sea, this suggests that the water mass is relatively closely associated with the depth range.

The results obtained here are in good qualitative agreement with the measurements of Packard et al. (1975) on electron transport activity in marine zooplankton. For epipelagic zooplankton, he obtained equivalent oxygen utilization rates at 18°C of .18 ml/l/y (Foraminifera, Copepods; his experiment No. 17) and 0.61 ml/l/y (Copepods, his experiment No. 15). For 2000 m to 3000 m, he obtained a rate of .0004 ml/l/y for a bathypelagic zooplankton assemblage (Foraminifera, Pteropods and Copepods, his experiment No. 16). This is consistent with our extrapolated (2500 m) deep water estimate of .0005 ml/l/y. The activation energies obtained by these workers (~ 15 k cal/mole, op cit.) are inadequate to explain the magnitude of the observed AOUR-depth trend on the basis of temperature change alone. Further, the pressure dependence of respiration is too weak (King and Packard, 1975). This trend must therefore be attributed to other processes, such as variation in available food or gradation in zooplanktonic speciation with depth.

8. Summary and conclusions

It appears that although a purely vertical exchange model can be constructed to explain the present (1977) vertical distribution of tritium in the upper 1 km of the Sargasso Sea, this approach fails in a number of ways. First, the vertical model cannot be satisfactorily extrapolated back in time to fit past tritium data. Second, the vertical diffusivities required are unrealistically large compared to estimates based both on thermocline structure and other geochemical tracers. Third, the presence of a mid-depth maximum in tritium indicates substantial lateral transport effects. (While this does not invalidate the model as a whole, it certainly limits its applicability to the upper 1 km). Fourth, the vertical model cannot simultaneously satisfy both the tritium and its daughter, ³He, distributions. This last failure is perhaps the most severe shortcoming and results from the effective complementary nature of their boundary conditions. An upper limit of .1 cm²/sec may be placed on the vertical diffusivity from ³He flux balance considerations.
A simpler, but physically more realistic model can be constructed based on lateral transport whereby the interior is ventilated via in-mixing or advection of late winter mixed layer water. The premise of internal homogeneity is supported not only from the latitudinal distribution of tritium, but also on the fact that geostrophic estimates indicate that recirculation occurs on timescales much shorter than the ventilation timescales. The ventilation timescales predicted by tritium and by $^3$He are mutually consistent, indicating that the lateral transport model (LTM) may indeed be a realistic representation of the processes responsible for ventilation of the thermocline. In addition, the past evolution of the tritium distribution is very well predicted by the LTM, despite the fact that it was not “tuned” to do so. The LTM appears quantitatively successful.

The vertical diagnostic model requires vertical diffusivities at least an order of magnitude greater than that predicted from the $^3$He flux balances and many other estimates. This suggests that less than ten percent of transport in the thermocline is by vertical processes. The elimination of vertical turbulent diffusion as an important transport process in the thermocline limits the type of thermocline models applicable to the oceans.

A comparison with Worthington’s residence time estimates is consistent with a more advective renewal in deeper waters (as opposed to a relatively diffusive exchange), but there appears to be a significant discrepancy in the lower thermocline, where the deep tritium maximum occurs. Worthington’s budgets do not account for this, and a substantial influx (as much as 2.5 Sv) would be required to achieve the ventilation observed, if it occurred by direct advection. However, a simple scheme of tagging the anticyclonic flow by eddy exchange under the Gulf Stream extension with northern, high tritium waters could explain this feature.

Finally, it is possible within the framework of the LTM to estimate the apparent oxygen utilization rate (AOUR). As might be expected, the AOUR decreases monotonically with depth, with near-surface rates of the order of 0.5 ml/l/y decreasing to rates of the order of .005 ml/l/y at 1800 m with an e-folding depth of about 350 m. Riley’s earlier AOUR estimates agree rather well with this work in shallow strata, but diverge to lower values in deeper water, a fact which primarily results from his using a “poorly ventilated ocean” model. Further, Packard’s E.T.S. results are consistent with the values observed. The oxygen minimum layer appears to be generally a result of a decreasing ventilation rate coupled with a decreasing consumption rate, and its presence may be enhanced by the influx of relatively “new” water under it (as evidenced by the deep tritium maximum).

In summary, it can be concluded that the processes primarily responsible for the ventilation of the Sargasso Sea main thermocline are isopycnal. Although the scope of the tritium and $^3$He data to date do not warrant an extrapolation to subtropical gyres in general, the work of Gregg (1977), Gargett (1976) and others suggest this may be a general conclusion. The geochemical ramifications of this observation are
important. The aliasing of the present distribution of tritium (or other tracers) by physically unrealistic models leads to incorrect estimates of the fluxes of other anthropogenic substances into the oceans, especially where the temporal variation of the input of these substances varies in a fashion different to tritium as, for example, in the case of CO$_2$ or the halomethanes. Perhaps more importantly, incorporation of unrealistic mechanisms will lead to incorrect predictions of what future fluxes will be, not only under present climatic conditions, but also under variations in climate forcing, whether natural or anthropogenically precipitated (as with the fossil fuel carbon dioxide problem). This paper represents an attempt at resolving some of the fundamental processes involved in oceanic transport.

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