On the climate of a subtropical ocean gyre: Decade timescale variations in water mass renewal in the Sargasso Sea

by William J. Jenkins[1]

ABSTRACT

A simple concept model of the climatology of a subtropical ocean gyre is developed on the basis of assumed isopycnal transport and latent heat flux. The 27 year record of annually averaged salinities on isopycnals near Bermuda shows clear and systematic variations, and significant (> 95% confidence) correlations with oxygen (positive) and vertical density gradient (negative) support the hypothesis. ^3He variations at two time periods are also consistent. Comparison with the Marsden Square 115 heat flux record shows moderately good lag correlations with increasing lag with depth. The seasonal records show the post winter arrival at the station of renewed water. Back calculation of renewal rates based on oxygen, salinity and ^3H-^3He data shows a factor of two variation in shallow water mass renewal rates over the past three decades.

1. Introduction

The oceans transport a significant amount of heat on a global scale (e.g., Ellis *et al.*, 1978; Bryden and Hall, 1980) and have a large thermal inertia. They therefore play an important role in the determination of both local and global climate, on timescales ranging from diurnal to centuries. The processes by which the oceans gain and lose heat, and by which this heat is transported, are themselves governed by climate, so that any realistic climate models must incorporate to some degree this coupling. The spectrum of atmospheric climatic variance is red, and recognition and measurement of an oceanic response (i.e., "ocean climate") is an important step in the direction of characterizing and understanding ocean-atmosphere coupling.

Whereas we have some knowledge of ocean climate for very long scales (from paleontological evidence, e.g., Streeter and Shackleton, 1979; from ^{14}C dating of the abyss, e.g., Munk, 1966 and Craig, 1969), and for very short timescales (e.g., mixed layer microstructure, Gargett *et al.*, 1979), the very critical annual to century timescales are not well determined. It is these intermediate timescales which are important for matters such as the impact of anthropogenic substances on the ocean, and in particular the fossil fuel CO_2 problem. The latter is of interest in that CO_2 is

1. Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, 02543, U.S.A.

a potential climate modifier and hence is coupled in a complex way to the processes which are responsible for the transfer (and ultimate demise) of this CO_2 into the oceans: namely, water mass formation and the general circulation.

Observation of the penetration of man-made substances into the oceans is supplying information about oceanic transport processes which occur on decade timescales. Nuclear weapons-produced isotopes such as tritium are providing useful constraints on mixing and advection (e.g., see Rooth and Ostlund, 1972; Jenkins and Rhines, 1980; Fine and Ostlund, 1980) especially when coupled with its stable, inert daughter ^{3}He (Jenkins and Clarke, 1976; Jenkins, 1980). Regardless of the tracer used, however, some caution must be exercised in interpreting tracer measurements characterized by timescales comparable to the timescales of variability in the transport processes themselves.

Worthington has discussed the role of heat flux in watermass formation (1968, 1972), and response of watermass formation and circulation to variations in climate forcing (1972, 1977). Bunker (1976, 1980) and Bunker and Worthington (1976) have used a large data base of ship weather reports to construct air-sea heat exchange charts and time series for the North Atlantic. What needs to be done now is to couple these heat flux measurements with a time series measurement of watermass formation, and thereby provide a means of testing, using and quantifying the relationships suggested by Worthington. This paper is an attempt to document an approach for observing and quantifying decade timescale variations in the rates of watermass formation in a subtropical gyre (the Sargasso Sea) and to link this variation with air-sea heat fluxes.

2. Background

Whereas a large number of studies have been made of the variations in space and time of sea-surface temperature, considerably less have been done on subsurface variability, largely due to the dearth of time series hydrographic data. Schroeder *et al.* (1959) examined historical soundings in addition to a then short timeseries of observations near Bermuda, in an attempt to characterize variations in Eighteen Degree Water, but the quality and frequency of data was too low to draw significant conclusions. Barrett (1969) noted a salinity change in the abyssal Sargasso Sea, but again the signal was not strong enough to make a conclusive climate link. Pocklington (1972), using the *Panulirus* timeseries data near Bermuda, also noted a general trend in salinity, temperature and σ_θ in the top 400 m. The use of heavy time averaging (5 year running mean) to some extent limits the utility of comparison to climate trends for such a relatively short series (15 years), but the general trends were at least consistent with one another. More recently, Frankignoul (1981) has examined the temperature spectrum for the *Panulirus* data (1954 to 1978) and, among other things, noted a significant difference in mean temperatures between the first and second halves of the 24-year record.

Recent studies have shown evidence of the impact of anomalous meteorological conditions on what is assumed to be "normal ocean" conditions: for example due to the especially severe winter of 1976-1977 (Diaz and Quayle, 1978; Worthington, 1977; Leetmaa, 1977). In particular, McCartney *et al.* (1980) reported the ubiquitous presence of anomalous lenses of both high and low salinity waters. The latter they attributed to large convective formation events in the Labrador Sea. Lazier (1980) reported evidence from OWS Bravo of strong secular variability in water-mass renewal in the Labrador Sea.

3. Approach

The data used in this study consist of a 27-year series (1954-1980) of almost 500 hydrographic soundings taken from the R.V. *Panulirus* near 32°10′N, 64°30′W in the Sargasso Sea. From the viewpoint of observing interannual variations in the regional scale water masses, this represents an ideal site since it is in the recirculation region of the Sargasso Sea gyre. The relatively rapid recirculation (Worthington, 1976) compared to the ventilation timescales (Jenkins, 1980), particularly in the upper thermocline, ensure that for timescales in excess of a year or two, the station is representative of water mass conditions in the gyre as a whole.

The station was occupied on the average about 1.5 times per month so that aliasing of seasonal and longer timescales is not a serious concern. Figure 1(a) shows the annual sampling rate throughout the record, and although there are gaps, and a slight decreasing trend, there appears to be more than adequate sampling density for timescales in excess of one year. Figure 1(b) shows the seasonal sampling bias, as given by the ratio of the number of observations in the winter (November-March) to the number in the entire year. A completely unbiased year would correspond to a ratio of 0.50. Although there appears to be a slight "fair weather" bias, there does not appear to be any substantial systematic deviations, and as may be seen later, no strong correlation with the observed property variations.

The approach to be used here is to compute property averages on isopycnal levels rather than at standard depths. The major rationale for this approach is threefold. First, it is proposed that the bulk of transport occurs along isopycnal surfaces rather than across them (Iselin, 1939; Montgomery, 1938) so that an isopycnal framework represents a more natural coordinate system. This concept has been shown to be consistent with observations of transient tracers in the Sargasso Sea (Jenkins, 1980). Second, a large component of "dynamical noise" associated with vertical displacements of isopycnals is eliminated. In particular, there exists a large eddy peak in the temperature spectrum (Wunsch, 1972) which would contaminate the lower frequencies. Third, since isopycnal "outcrops" move with the season, a substantial part of seasonal variability associated with such movements (and hence possible aliasing of the decade timescale record) is removed in this coordinate system. This last point will be discussed further, later in this paper.

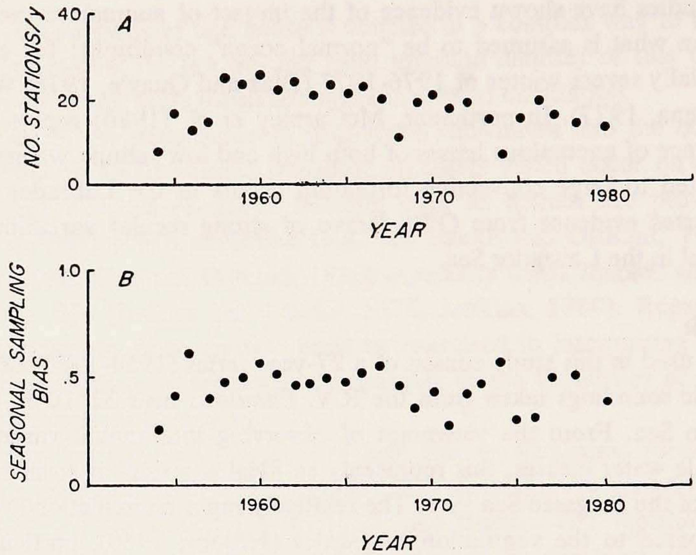

Figure 1. (a) Sample rate (stations per year) and (b) sampling bias (fraction of winter sampling) throughout the *Panulirus* timeseries (1954-1980).

The physical model envisaged in this study is a wind-driven gyre consisting of isopycnal layers that outcrop at the sea surface within or very near the gyre, and which are "fed" by Ekman pumping at the sea surface (cf. Stommel, 1979). The primary impedance against mass transfer or exchange is pictured as being in the regions of outcropping rather than within the interior, at least for the shallowest layers. Because mass must be conserved, in the absence of diapycnal buoyancy modification we must require eventual "leakage" or direct flow of this surface modified water out of the gyre. In this respect this is an open-ended box model as was used by Jenkins (1980) for explaining ^{3}H, ^{3}He distributions in the Sargasso Sea. Clearly, this simple concept cannot hold below those density levels that characteristically outcrop within the high velocity core of the gyre. For example, surfaces which outcrop within the slower, more easterly part of the gyre (e.g., $\sigma_\theta \sim 26.8$) will show gradients due to lower velocities. The deeper waters, however, apparently "outcrop" at a front corresponding to the northern edge of the gyre. This can be seen quite dramatically in the meridional distributions of tritium (Ostlund *et al.*, 1974) and ^{14}C (Stuiver and Ostlund, 1980) where there is an abrupt decrease in those isotopes at about 40N. This transition was also noted for oxygen by Worthington (1962, 1976) and indeed constituted a fundamental argument for his two gyre hypothesis. It is not clear how the "permeability" of this boundary is controlled or how it varies with time, but McCartney *et al.* (1980) have observed penetrations of lenses of anomalously fresh water along 55W following the winter of 1976-1977,

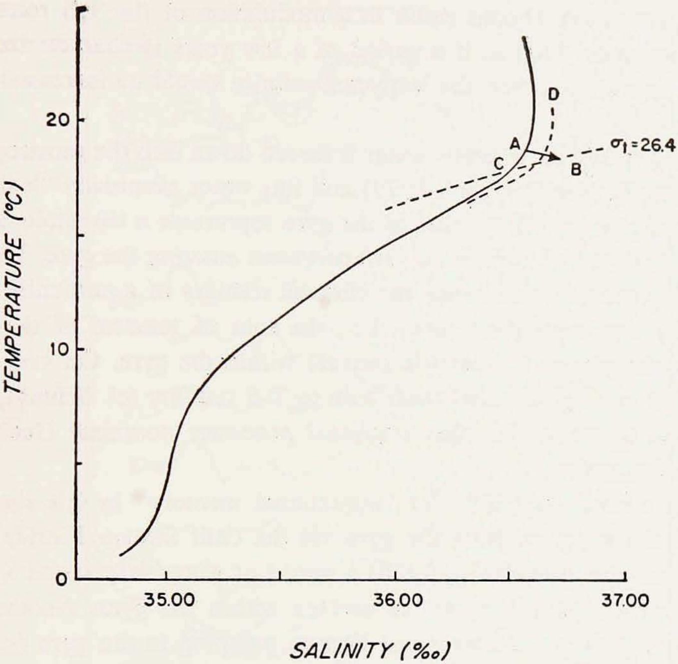

Figure 2. A schematic representation of what occurs during heat loss from the Sargasso Sea. Largely latent heat flux results in a net increase in salinity as well as cooling (the vector A-B). Isopycnal mixing results in exchange between waters B and C, tending to draw the T-S relation toward a curve such as D. Years of larger than normal heat flux result in a "pulling" of the T-S relation more toward D, resulting in higher salinities on isopycnal surfaces.

an artifact presumably of the severe meteorological conditions of that winter (cf. Diaz and Quayle, 1978).

The focal point of this analysis will be the salinity record, as averaged annually, on isopycnals. A major portion of heat transfer from the ocean to the atmosphere in this area is latent heat of evaporation (e.g., Bunker, 1975, 1976). Consequently, as water loses heat, it increases in salinity. Mixing and transport of this water occurs along an isopycnal, so that heat loss from the ocean increases salinity (and counter-intuitively, temperature) on an isopycnal. Figure 2, in a cartoon form, shows this process on a T-S diagram. For example, water characterized by temperature and salinity "A" is cooled by largely latent heat transfer at the surface and its salinity is correspondingly increased by evaporation so that its T-S properties progress along the vector to "B". Mixing along isopycnals with water type "C" tends to pull the T-S relation out toward higher salinities. Although this heat loss mechanism must act in concert with other processes to maintain the approximate T-S relation for the region (i.e., in some quasi-steady state), variation in climatic forcing (i.e., in the net annual latent heat flux), should result in a modulation of the T-S relation and of salinity on isopycnals. That is, if a period of a few years is characterized by more heat removal than some other, the isopycnal salinity would be increased more than in other years.

If one pictures a system whereby water is forced down into the geostrophic regime by Ekman pumping (see Stommel, 1979) and this water eventually "leaked" out of the gyre, then the mean T-S relation of the gyre represents a climatological average of the meteorological T-S imprint on surface waters entering the gyre. The timescale of the averaging process, and hence the climatic stability of a particular water type is governed by its "ventilation rate," i.e., the rate of renewal of the watermass divided by its abundance (volumetric census) within the gyre. On very long timescales diapycnal mixing may also contribute to T-S stability (cf. Schmitt, 1981) but it appears from tracer studies that isopycnal processes dominate (Jenkins, 1980) on the decade timescales.

There is, in a sense, potential for "interannual memory" in this simple model. Since most of the water entering the gyre via the Gulf Stream/Florida Current is recirculated from the gyre (Iselin, 1936) a parcel of water may spend a number of years (i.e., several revolutions) at the surface within the gyre. Another point of comparison is that the total downward Ekman pumping in the gyre ($< 10^7 m^3 S^{-1}$, Leetmaa and Bunker, 1978) is small compared to the recirculation ($\sim 4 \times 10^7 m^3 S^{-1}$ for $T > 17°C$, Worthington, 1976). Again, this suggests that surface water may in fact spend several years in the gyre, and that a buildup of warm saline cap-layer from previous winters may precondition the following year's renewal. Thus, some lag may occur between the meteorological forcing and the oceanic "response." Characteristically, this lag might be of the order of the circulation timescale of the upper part of the gyre: i.e., a few years.

4. The isopycnal salinity record

The data were block-averaged by year and potential density interval ($\pm$ .05‰ in σ_θ) and then quadratically interpolated against σ_θ to the standard σ_θ levels. In all cases the interpolation range was small (generally less than 0.01‰ in σ_θ), and resulted in small differences between the block average and the interpolated value. The only filtering applied to the data was to reject any data points more than 3σ from the mean for each density interval. This procedure resulted in the rejection of less than 1% of the data.

Figure 3 shows the time trends of salinity on isopycnals for $26.1 \leqslant \sigma_\theta \leqslant 26.8$‰ from 1954 to 1980. There appears a clear variation in the isopycnal salinity on all surfaces down to $\sigma_\theta = 26.8$ or about 550 m depth. The salinity variations in this upper range are quite coherent, with the deeper layers ($\sigma_\theta = 26.5$-26.7) lagging the upper layers by not more than one to two years. The magnitude of the variations is

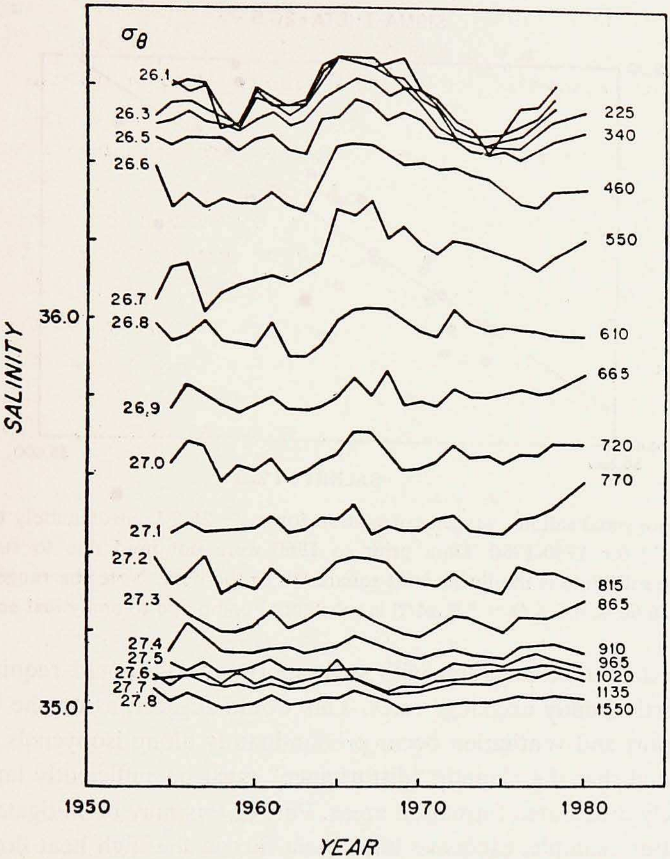

Figure 3. The timeseries of mean annual salinities on isopycnals ($26.1 < \sigma_\theta < 27.8$). Approximate depths for the isopycnals are listed on the right. Note the large (amplitude > 0.1‰) coherent decade timescale variations for $\sigma_\theta < 26.7$.

large, being about 0.2‰ in salinity, and corresponds to temperature variations as large as .5°C; i.e., more than an order of magnitude greater than analytical error, and much larger than variations in the deeper ($\sigma_\theta > 27.2$) strata. The variations that are seen in the deeper strata appear more random and may be related to sampling and standardization errors. Mantyla (1980) has compared salinity standard water batches for post 1959, and the largest deviations are less than .01‰ so that it appears unlikely that the deepest salinity variations can be attributed solely to standardization error. What is clear, however, is that the shallow salinity variations are substantial compared to those of the deeper strata.

The apparent propagation of this "isopycnal salinity" (IS) variation to 500 or 600 m depth is remarkable in that significant local ventilation only occurs to $\sigma_\theta \sim$

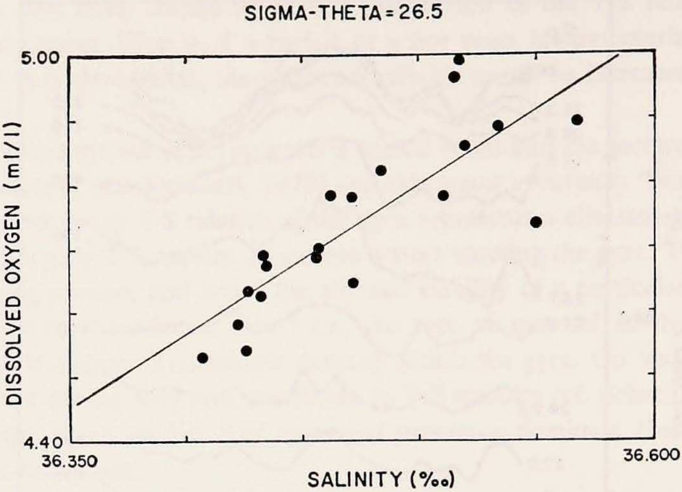

Figure 4. The isopycnal salinity-oxygen correlation for $\sigma_\theta = 26.5$ (approximately the "Eighteen Degree Water") for 1960-1980. Data prior to 1960 were not used due to standardization problems. The solid line is the linear least-squares fit to the data. Note the range in variation ($36.40 < S < 36.60$‰, $4.5 < O_2 < 5.0$ ml/l) is substantial compared to analytical accuracies.

26.5, and the deeper propagation with such short timelag would require unrealistically high vertical eddy exchange rates. This would suggest, as do the tracer data, that the transport and ventilation occur predominantly along isopycnals rather than across them, and that the climatic "disturbance" must be sufficiently large scale to affect the widely distributed formation areas. Part of this may be mitigated by ocean surface flow. For example, excessive latent heat flux in the high heat flow region in the northwest Sargasso Sea (see Bunker and Worthington, 1976) would result in positive salinity anomalies that would be advected eastward into the formation regions for denser water types.

The hypothesis that the observed positive IS variations are associated with greater than normal heat flux/water mass formation, requires a positive correlation between the dissolved oxygen and the IS. That is, although the observed dissolved oxygen content is in approximate balance between respiration and ventilation, variation in the latter process will result in isopycnal oxygen anomalies. Oxygen measurements at this and other stations prior to 1961 suffered from standardization problems (Worthington, private communication), making the oxygen record before 1961 unusable for this purpose. However, for the 20-year record after 1960, the data show a very clear trend. Figure 4 shows an example of this positive correlation (for $\sigma_\theta = 26.5$, corresponding approximately to the 18° water). The linear correlation coefficient r, shown in Figure 5 as a function of σ_θ shows a broad positive maximum, generally exceeding 95% confidence between $\sigma_\theta = 26.3$ and $\sigma_\theta = 27.0$ and often (e.g., $\sigma_\theta = 26.5$ to 26.7) exceeding 99.9% confidence.

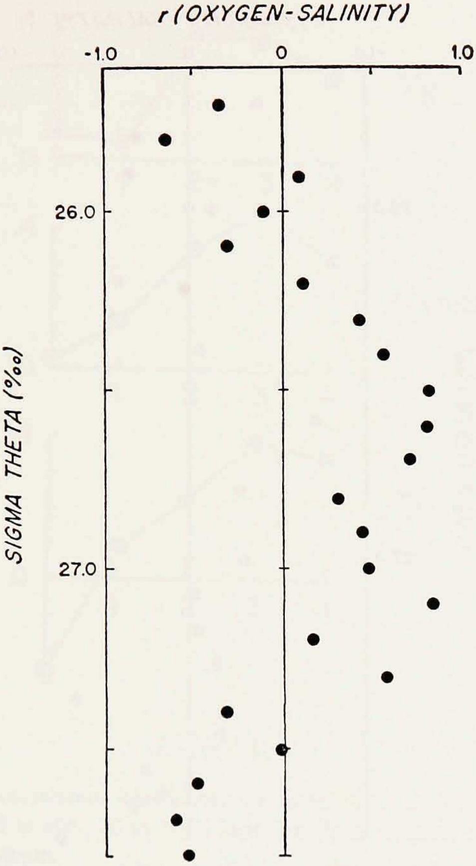

Figure 5. The isopycnal salinity-oxygen regression coefficient as a function of σ_θ. Note the broad and significant positive maximum centered around $\sigma_\theta = 26.5$ and also the trend to negative correlation in deeper strata. The former results from the latent-heat-ventilation correlation described in the text, while the latter indicates subarctic mode influence.

Shallower strata show a weak, negative correlation. The weakness of the correlation is presumably due to the dominance of local vertical processes in sustaining oxygen concentrations. The negative trend of the correlation may in part be due to the inherent negative dependence of oxygen solubility on salinity and temperature, but the statistics here are inadequate to establish this. The trend in deeper strata toward a negative IS-O_2 correlation may be interpreted as a transition toward subarctic modes of water mass formation as characterized by an excess of precipitation over evaporation. That is, the penetration of high oxygen, low salinity waters from the subarctic regions will produce a negative salinity-oxygen correlation. Within the framework of the simple concept model posed here, one would picture climatological variations in the permeability of the northern "wall" which would be

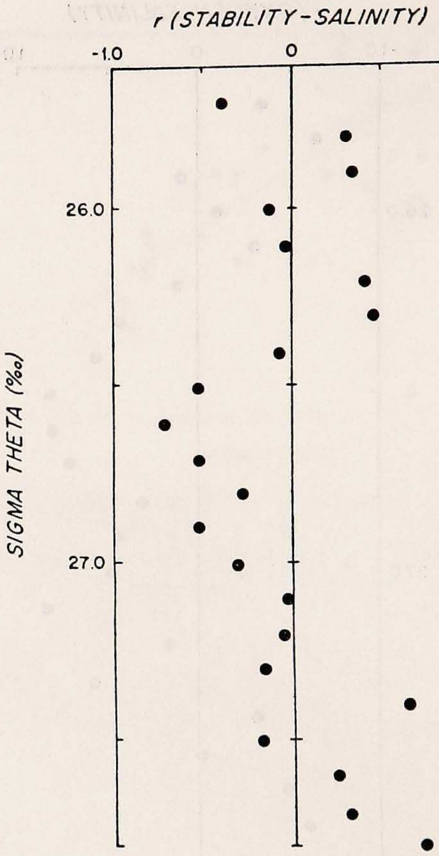

Figure 6. The isopycnal salinity-stability correlation coefficient *vs.* σ_θ. The data are much more scattered than Figure 5 due to the noise levels introduced by computing density gradients from discrete hydrographic data. Note, however, the broad trend toward a negative extremum around $\sigma_\theta = 26.5$ to 26.7.

marked by variable and perhaps sporadic incursions of high oxygen, low salinity water. The presence of low salinity lenses of water at 55N in 1977 (McCartney *et al.*, 1980) is suggestive of this influence as well.

A second effect expected on the basis of this simple model is an anti-correlation between the vertical stability, as given by $\frac{d\rho}{dz}$, and IS, in the region of the Subtropical Mode ("Eighteen Degree") Water. That is, more copious production of the upper thermocline mode waters would be expected to result in a weaker vertical density gradient (due to the "thermostad" effect). Since one is dealing with a first derivative property the correlation will be noisier, but the broad negative trend seen in Figure 6 is as expected. The general trend toward a positive correlation in deeper waters is possibly a weak compensation for the upper trends (a strengthening of the

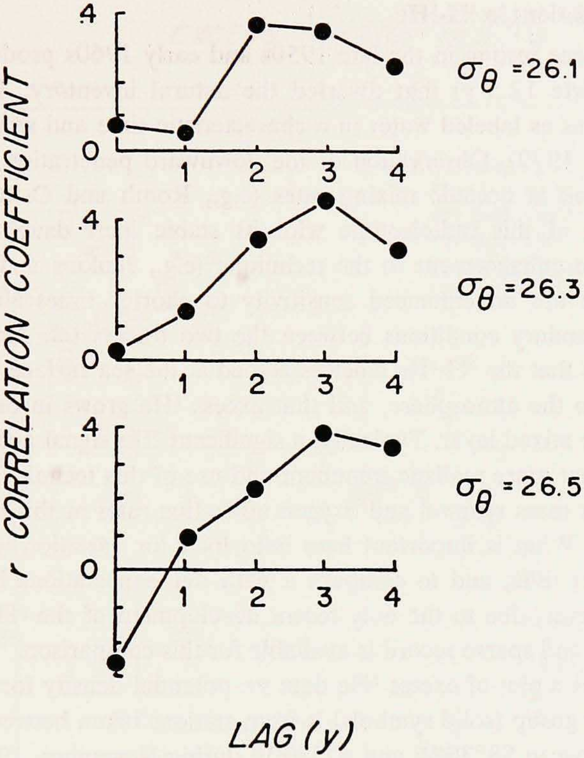

Figure 7. Lagged linear correlation coefficients on three surfaces for isopycnal salinity *vs.* Marsden Square 115 (30 to 40N, 60 to 70W) heat flux (from Bunker, 1975). Note the apparent increase in lag with depth.

density gradient in the lower thermocline in response to a weakening in the upper thermocline), but further analysis is required.

It appears then, that significant IS variations are seen in the shallowest strata ($\sigma_\theta < 26.7$) and that these variations correlate with oxygen and stability in a way consistent with the simple concept model presented. It then remains to compare the IS record with annual average areal heat flux. An appropriate measure is Bunker's (1975) average annual heat flux for Marsden Square 115. Although this data extends only until 1972, it represents a good areal average of a large number of ship observations. Figure 7 shows the lagged linear correlation coefficient between the heat flux and IS for the three surfaces (σ_θ = 26.1, 26.3 and 26.5) showing the increasing lag with depth. The correlations are only good at the 90% confidence limit due to the short record, but the systematic progression to increasing lag with depth is suggestive. The lag times in excess of one year require some means of maintaining a "memory" of previous years, and are consistent with the arguments presented in Section 3 that the gyre "stores" high salinity water in the surface layers.

5. Temporal variations in ^{3}H-^{3}He

Nuclear weapons testing in the late 1950s and early 1960s produced an amount of tritium (half-life 12.5 y) that dwarfed the natural inventory. This tritium has entered the oceans as labeled water in a characteristic time and space pattern (e.g., see Weiss *et al.*, 1979). Observation of the downward penetration of this dye has been used to look at oceanic mixing rates (e.g., Rooth and Ostlund, 1972), but the combination of this radioisotope with its stable inert daughter ^{3}He is providing a valuable enhancement to the technique (e.g., Jenkins and Clarke, 1976). The gains made are an enhanced sensitivity to shorter timescales and complementarity of boundary conditions between the two tracers (cf. Jenkins, 1980). A simplistic view is that the ^{3}H-^{3}He clock is zeroed at the sea surface by escape of all tritiugenic ^{3}He to the atmosphere, and that excess ^{3}He grows in once the water is isolated from the mixed layer. Typically, a significant ^{3}He signal accrues in a month or so. A somewhat more realistic treatment and use of this technique has been used to estimate water mass renewal and oxygen utilization rates at this location in 1977 (Jenkins, 1980). What is important here is to look for variation in this measured renewal rate over time, and to compare it with our expectations based on the IS variations. However, due to the only recent development of the ^{3}H-^{3}He technique, only a very brief and sparse record is available for this comparison.

Figure 8 shows a plot of excess ^{3}He data *vs.* potential density for a few points in time. The earlier group (solid symbols) is from stations taken between 30°30′N and 34°30′N and between 58°30′W and 67°30′W during November, 1974, and March, 1975. The later group (open symbols) consists of profiles taken at the *Panulirus* station from March, 1977 to November, 1978. Although the time separation between the groups is small, there appears a marked difference in the ^{3}He values for the two data sets shallower than $\sigma_\theta = 26.6$. This is expected on the basis of the salinity trends in Figure 3. The earlier record has substantially higher ^{3}He excesses (and therefore poorer ventilation) for density surfaces between 26.2 and 26.5.

The magnitude of this difference is, for example, equivalent to a six-fold reduction in ^{3}He standing crop for $\sigma_\theta = 26.3$. That is to say, $\sigma_\theta = 26.3$ was very effectively degassed or ventilated in the intervening period. The observed ^{3}He standing crop excess (relative to 1977) observed in 1974-1975 was equivalent to more than three years' decay of tritium—suggesting again a mechanism for interannual memory, and one which appears to respond rapidly to events of severe meteorological stress, but relaxes slowly under years of poor ventilation.

In fact, the marked difference observed in ^{3}He is reflected in oxygen, salinity and vertical stability in precisely the way predicted by the simple concept model presented here (Table 1). The change is consistent with the particularly severe winter of 1976-1977 (e.g., see Diaz and Quayle, 1978) and appears related to the proposed Gulf Stream intensification reported by Worthington (1977).

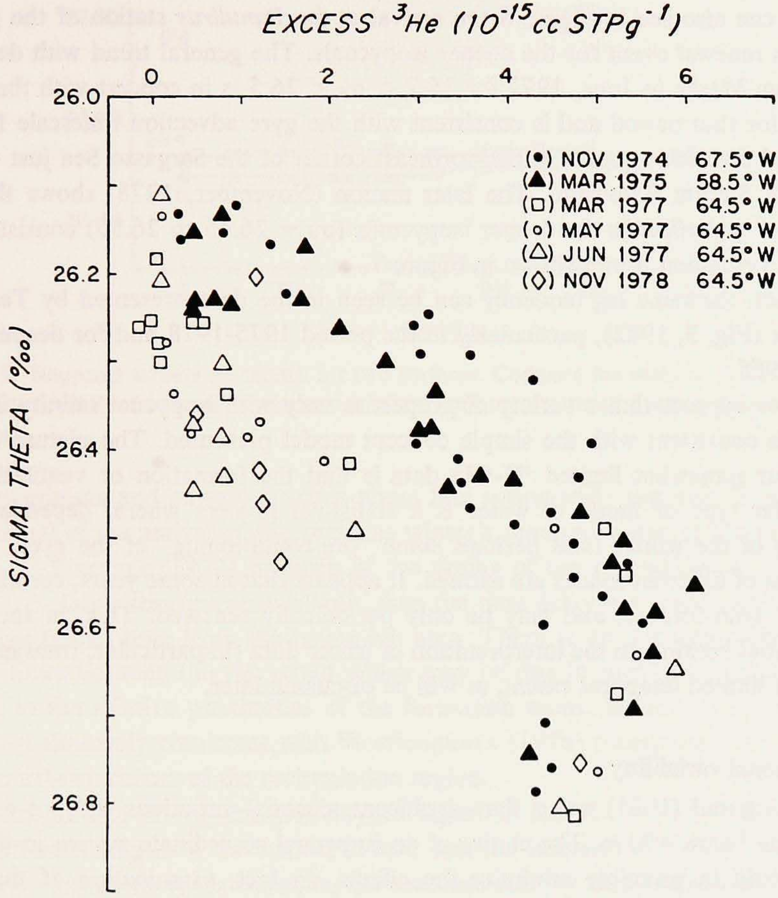

Figure 8. Excess ^{3}He *vs.* density at *Panulirus* for two periods in time (1974-1975 and 1977-1978). The open symbols correspond to the later period. The envelopes for the two data sets separate between $\sigma_\theta = 26.1$ and 26.5 and converge universely for $\sigma_\theta > 26.5$. The rationale is that the amount of excess ^{3}He is inversely related to the degree of ventilation, so that 1974-75 is a period of poor ventilation relative to 1977-1978, as would be predicted from Figure 3.

Table 1. Observed property differences between 1974-1975 and 1977-78.

σ_θ	ΔS(‰)	ΔO_2(ml/l)	$\Delta\rho_z$(10^{-5}g cm^{-4})	Δ^3He*
26.2	.036	.31	—	−1.3
26.3	.089	.22	−.52	−2.5
26.4	.068	.22	−.76	−2.0
26.5	.030	.10	−.26	−0.5

* Units 10^{-15}cc(STP)g^{-1}, one unit corresponds to about 1.2 y isolation time.

One can also see in Figure 8 the arrival at the *Panulirus* station of the previous winter's renewal event for the deeper isopycnals. The general trend with decreasing ^3He from March to June, 1977 for $26.2 < \sigma_\theta < 26.5$ is in concert with the oxygen record for that period and is consistent with the gyre advection timescale from the expected formation region in the northeast corner of the Sargasso Sea just south of the Gulf Stream Extension. The later station (November, 1978) shows the even-more-delayed effect in the deeper isopycnals ($\sigma_\theta \sim 26.45$ to 26.55) consistent with the lag correlation trends shown in Figure 7.

In fact, the same lag tendency can be seen in the data presented by Talley and Raymer (Fig. 3, 1982), particularly in the period 1975-1978 and for deeper modes 1963-1965.

It thus appears that a variety of properties vary with isopycnal salinity in a way which is consistent with the simple concept model presented. The picture evolving from our somewhat limited ^3H-^3He data is that the formation or ventilation of a particular type or mode of water is a statistical process where, dependent upon severity of the winter (and perhaps some "pre-conditioning" of the gyre), varying amounts of different modes are formed. It appears that in some years, certain modes are not even formed, and may be only periodically renewed. This, in fact, has a substantial bearing on the interpretation of tracer data (in particular, transient tracer data) of limited temporal extent, as will be discussed later.

6. Seasonal variability

Frankignoul (1981) noted that significant seasonal variations did not appear to penetrate below 400 m. The choice of an isopycnal co-ordinate system in this analysis should in principle minimize the effects. In fact, examination of the month averaged IS data shows a much smaller seasonal amplitude than the interannual trends. Also, examination of the sampling record shows that there were no years of pronounced "favoring" of one particular season or month-group, so that likelihood of strong seasonal influence on the inter-annual record is small. However, because the seasonal signal is much smaller than the long-term variations, the reverse may not be the case. Consequently, only those years with complete 12-month records were included in constructing monthly means. Figure 9 shows the monthly trends in IS for $\sigma_\theta = 26.4$ and $26.5‰$. Compare the vertical scale (salinity) to Figure 3 for the interannual IS trends. The signal amplitudes are substantially smaller than the interannual variations justifying the initial approach of simple block averaging for calculation of the interannual trends.

The oxygen and stability $\frac{\delta \rho}{\delta z}$ records, however, show a significant seasonal signal. Figures 10 and 11 show data for $\sigma = 26.4$ and $26.5‰$ respectively (note the scale differences between the two figures). Both isopycnals show trends toward

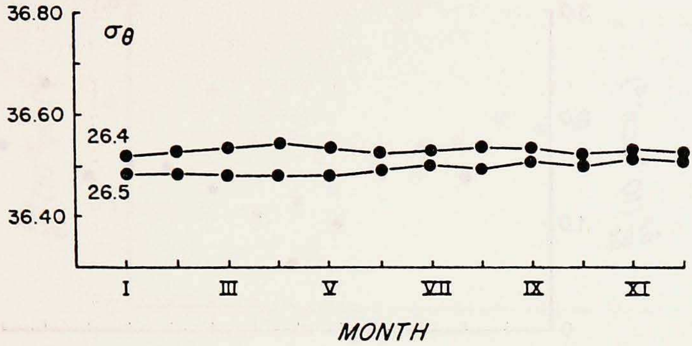

Figure 9. Isopycnal salinity *vs.* month for two surfaces. Compare the scale to that of Figure 3 and note that the "seasonal" variations on these surfaces are small compared to interannual trends.

stability minima and oxygen maxima about late spring-early summer. The extrema are a signal of the arrival of the previous winter's renewed water at the monitoring station. If Warren's (1972) scenario of the timing of the renewal events is appropriate, i. e., late February-early March, then the time delay is a manifestation of the advective transit time from the formation area. There is an apparent difference in arrival times between the two strata which may be due in part to baroclinicity and in part to the relative proximities of the formation areas. In any case, the transit time is qualitatively consistent with Worthington's (1976) transports and formation in the northeast corner of the recirculation region.

The magnitude of the seasonal oxygen signal is notably higher than would be expected on the basis of the salinity record and the interannual IS-O_2 correlations. On the 26.5‰ surface, for example, the seasonal salinity amplitude is about .02‰, which when taken with interannual IS-O_2 slope of 2.6 ml/1/‰ accounts for only 25% of the observed seasonal oxygen variation. This effect, however, can be rationalized in the following way. The apparent decrease of oxygen after injection into the water mass is due to two effects: conservative mixing of the anomalous water into the watermass (i.e., in a way much like salinity would behave) and *in situ* biological consumption of the oxygen. The latter process (i.e., the oxygen utilization rate, OUR) can in fact be estimated from transient tracers (Jenkins, 1980) and (as discussed later, cf. Table 2) is of the order of .16 ml/l/y for $\sigma_\theta =$ 26.5. Within uncertainties of the estimate of the OUR, the oxygen trends seen in Figures 10 and 11 are consistent.

The significance of the large seasonal signal in oxygen (about 50% of the interannual signal) is that some aliasing of the interannual isopycnal oxygen record due to seasonal biases (Fig. 1b) will contribute variance to the IS-O_2 correlations (e.g., Fig. 4). The observed bias ratio variability (about 20%—see Fig. 1b) would introduce about .05 ml/l for $\sigma_\theta = 26.5$, which would explain most of the residual variance.

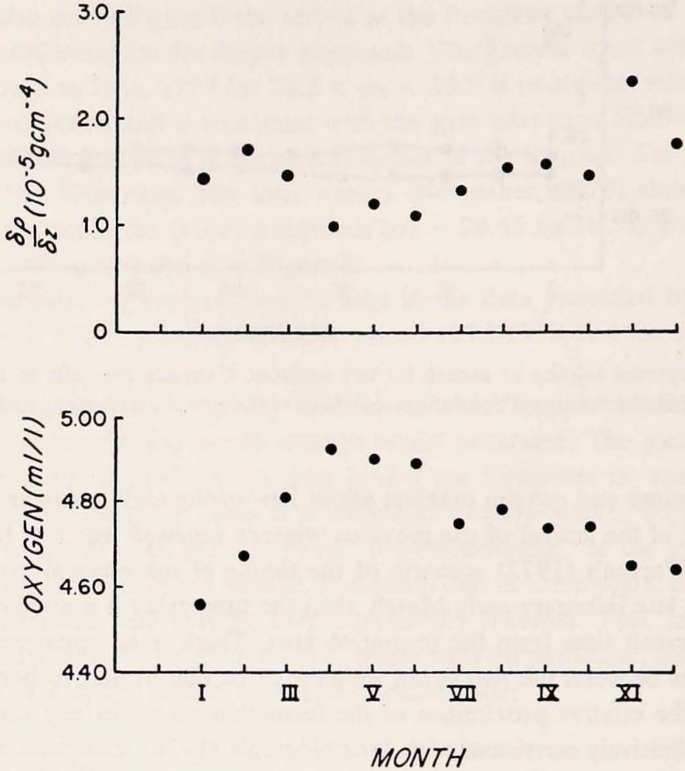

Figure 10. Seasonal (monthly averaged) vertical density gradient and dissolved oxygen for $\sigma_\theta = 26.4‰$. Note the broad minimum in ρ_z and broad maximum in dissolved oxygen at about April-June.

Table 2. Oxygen utilization rates and IS-AOU relations.

		IS-AOU Correlation Parameters			
σ_θ	OUR*	*a*	*b*	*r*	*n*
26.3	0.225	1.92 ± 0.59	−1.07 ± 0.38	−0.57	19
26.4	0.189	2.99 ± 0.75	−1.69 ± 0.49	−0.64	19
26.5	0.156	4.49 ± 0.58	−2.64 ± 0.39	−0.85	19
26.6	0.133	4.28 ± 0.51	−2.53 ± 0.38	−0.85	19
26.7	0.112	3.62 ± 0.48	−2.03 ± 0.40	−0.77	19
26.8	0.094	2.91 ± 0.75	−1.41 ± 0.76	−0.41	19
26.9	0.079	3.25 ± 0.52	−1.76 ± 0.64	−0.56	19

* Estimated by ^3H-^3He and regressed as in text (cf. from Jenkins, 1980).

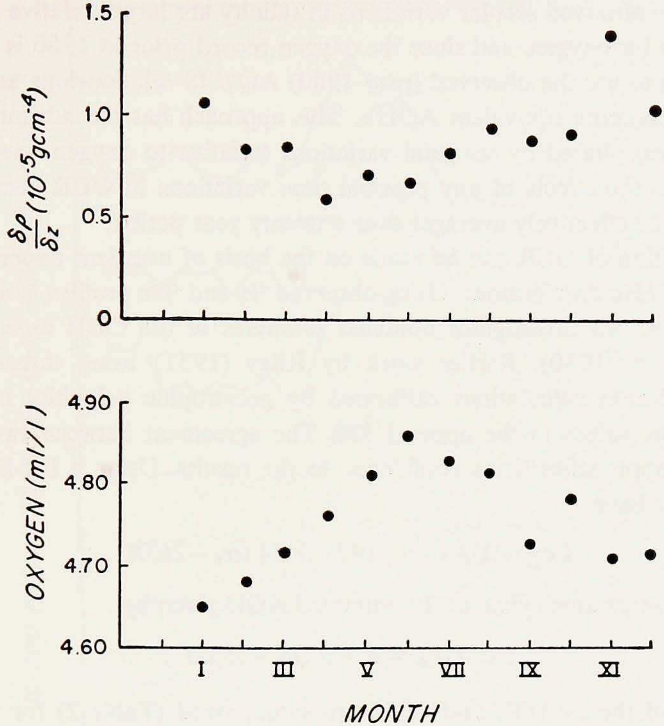

Figure 11. As in Figure 10, but for $\sigma_\theta = 26.5‰$. Note the differences in scale and the somewhat later extrema. The deeper stratum shows smaller changes as would be expected on the basis of ventilation rates.

7. The past record of water-mass formation

In principle, we now have a means of back-calculating the degree of ventilation or renewal of water on an isopycnal as a function of time. This may be achieved by combining the observed apparent oxygen utilization (AOU) record with estimated oxygen utilization rates (OUR). That is, we use biological respiration as a clock. The assumption is that the AOU record is controlled largely by the degree of ventilation, and that the OUR is to a first order time invariant. The latter is, to some extent, borne out in the IS-O_2 data: no significant curvature is seen in the IS-O_2 relations. Although not a strong test, it suggests that if some variation in OUR does occur, it is not a severe problem. The former is more difficult to argue, but on the basis of observed lateral gradients in the salinity field (e.g., see Worthington, 1976, Fig. 28 for the 15°C surface which corresponds roughly to $\sigma_\theta =$ 26.8), large lateral movements and/or fluxes (almost basin scale) would be needed to achieve the observed modulations.

Because the observed secular variations in salinity are larger relative to analytical accuracy than for oxygen, and since the oxygen record prior to 1960 is not reliable, I have chosen to use the observed (post 1960) AOU-IS relationships and the entire IS record to compute equivalent AOU's. This approach has the advantage that the IS record is less aliased by seasonal variations (relative to oxygen—see Section 5) and minimizes the effects of any possible time variations in OUR because the IS-AOU regression effectively averages over a twenty year period.

The estimation of OUR can be made on the basis of transient tracers: in particular, using ^3H-^3He distributions. Using observed ^3H and ^3He profiles in an isopycnal-outcrop model, this investigator obtained estimates of the OUR as a function of density (Jenkins, 1980). Earlier work by Riley (1951) using three-dimensional advection diffusion calculations calibrated by geostrophic velocities obtained virtually the same values in the upper 1 km. The agreement between two essentially independent approaches lends confidence to the results. Using a log-linear best-fit to the data, we have

$$\mathrm{Log}\,(\mathrm{OUR}) = -0.97 - 1.74\,(\sigma_\theta - 26.0)\ .$$

The AOU-IS regressions yield an IS estimated AOU given by

$$(\mathrm{AOU})_{\mathrm{IS}} = a + b\,(\mathrm{IS} - 35‰)\ .$$

Estimates of the IS-AOU and OUR are summarized (Table 2) for several density strata.

The rate of ventilation on a given isopycnal is then simply estimated from

$$R = \frac{\mathrm{OUR}}{(\mathrm{AOU})_{\mathrm{IS}}} \qquad y^{-1}\ .$$

The trends for a number of isopycnals ($26.3 < \sigma_\theta < 26.9$) corresponding to an average depth range from about 200 m to 600 m are presented in Figure 12. Note that the ordinate is in logarithmic coordinates so that the amplitude of variation is easily compared to the mean ventilation rates for individual strata.

As suggested from the salinity record, there appears a broad maximum in ventilation in the mid to late 1960's. The magnitude of the effect decreases with depth, with some evidence of lagging (cf. Fig. 7). The remarkable feature of the mid 1960s climatic event is that the onset (watermass response) was approximately the same for the different strata; but the relaxation was slower for the deeper levels. The latter aspect is at least consistent with the concept of increasing response time (ventilation timescale) with depth. The former is inconsistent with our earlier arguments about "preconditioning" of the gyre, inasmuch as onset of watermass formation may have occurred simultaneously over a large region, where horizontal gradations in surface density produced a distribution of water types. Furthermore, severe cumulative meteorological stress (i.e., high heatflux for extended periods)

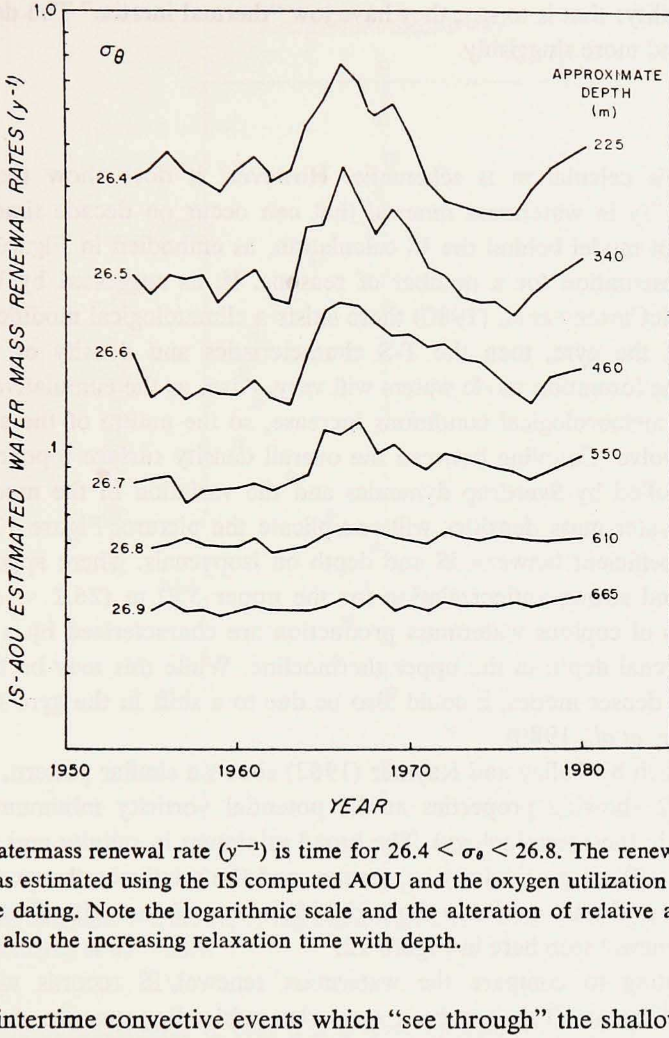

Figure 12. Watermass renewal rate (y^{-1}) is time for $26.4 < \sigma_\theta < 26.8$. The renewal (or ventilation) rate was estimated using the IS computed AOU and the oxygen utilization rates obtained from ^3H-^3He dating. Note the logarithmic scale and the alteration of relative amplitude with depth. Note also the increasing relaxation time with depth.

may force wintertime convective events which "see through" the shallower strata to the deeper layers, forcing the wintertime surface isotherms farther southward in a fashion similar to Stommel's Ekman Demon (Stommel, 1979), but on a larger scale. This would have the combined effect of producing simultaneous onset of renewal for a number of strata, and forcing renewal into deeper modes. This is consistent with the records shown here in Figure 12, and in the observed pycnostad density increase reported by Talley and Raymer for the critical mid 1960's period (1982, Fig. 5).

The relative scale of the variation in renewal rate is about a factor of two between "lean" and "fat" for the upper surfaces, and alternates downward. The trends observed are not unexpected, as the shorter residence time water masses (i.e., those which are ventilated on shorter timescales) should be more responsive to meteorological variability: that is to say, they have low "thermal inertia." The deepest water masses respond more sluggishly.

8. Discussion

Clearly, this calculation is schematic. However, it does show the magnitude of the variability in watermass renewal that can occur on decade timescales. The simple concept model behind the IS calculation, as embodied in Figure 2, is not a realistic representation for a number of reasons. If, as suggested by Worthington (1977) and McCartney *et al.* (1980) there exists a climatological modification of the circulation of the gyre, then the T-S characteristics and density of the "source waters" for the formation mode waters will vary. Also, as the cumulative departures from normal meteorological conditions increase, so the nature of the surface layer waters will evolve. Coupling between the overall density surface topography of the gyre as controlled by Sverdrup dynamics and the variation in the modal distributions of the water mass densities will complicate the picture. Figure 13 shows the correlation coefficient between IS and depth on isopycnals. There appears to be a very broad and strong anticorrelation for the upper 550 m ($26.1 < \sigma_\theta < 26.8$). That is, years of copious watermass production are characterized by a shoaling of average isopycnal depth in the upper thermocline. While this may be the filling of the gyre with denser modes, it could also be due to a shift in the gyre recirculation (cf. McCartney *et al.*, 1980).

The approach by Talley and Raymer (1982) shows a similar pattern, particularly their Figure 2 showing properties at the potential vorticity minimum (the "18° Water Core" in their terminology). The broad minimum in salinity and temperature in the early 1970's is particularly suggestive, and the relatively abrupt transition of the pycnostad to denser modes in the mid-1960's is consistent with the maximum in water mass renewal seen here in Figure 12.

It is tempting to compare the watermass renewal/IS records with external climatological factors. This can be a somewhat misleading practice due in part to the brevity of the record compared to the dominant timescale of variability, and due also to the fact that observed covariance does not necessarily prove causal relation, but may imply common cause. This last point, in fact, is important in that it may ultimately lead to an appropriate diagnostic model of the system. In the absence of a believable physical model of the water-mass renewal-climate forcing system, we must then content ourselves with simple comparisons between the observations and other external, but hopefully related parameters.

The covariance between the Bunker (1975) MS 115 heatflux (BHF) and the IS record is not as strong as we would like on the basis of our simple concept model. The lags observed (2-3 years, Fig. 7) border on the limits of the lag correlation techniques for the record lengths; a longer record is needed. Further, the choice of

Figure 13. The isopycnal salinity-depth regression coefficients *vs.* σ_θ. Note the broad negative maximum centered at $\sigma_\theta = 26.5$.

MS 115 is one of conscience in that it is a large enough area to have a statistically significant ship meteorological observation base, and includes the Gulf Stream and a large part of the associated high heat flux "sausage" described by Bunker and Worthington (1976). However, the buffering effect of the Bermuda High may damp the heat flux signal expected to be largest in the sausage. In addition, the choice of the annual average BHF is not necessarily the best, because the potential storage of density and salinity in the upper part of the gyre may mean that the cumulative departure from the mean is more appropriate. Also, the renewal response to climate forcing may be nonlinear.

Worthington (1968, 1971, 1977) argued that the impingement of cold, dry continental air on the Sargasso Sea gyre is responsible for a substantial part of the water mass modification and circulation. A signature of this phenomenon may be modulation of the frequency and path characteristic of cyclones. Hayden (1981) applied principal component analysis to the spatial pattern and time record of annual average frequencies of extratropical cyclones. The first eigenvector corresponded to the abundance of marine cyclones relative to continental cyclogenesis. More importantly, the second eigenvector of the analysis corresponded to the total coastal cyclogenesis, and centered along the coast at about 35N. The correlation between the time series of the second eigenvector weighting, and the $\sigma_\theta = 26.5$ IS record shows a moderately strong negative correlation ($r = -.49$, $P_{(\mathrm{corre.})} > .95$) which at first glance appears counter to our intuitions. However, the first eigenvector peaks in the early 1960's and reaches a minimum in the early 1970's, suggesting that there does indeed appear to be a marine cyclone maximum just prior to the mid-1960's IS maximum, and a minimum just prior to the mid-1970's IS minimum. The early-sixties marine cyclogenesis is maximum and early 1970's minimum is reflected by corresponding minimum and maximum east coast temperature records (cf. Mock and Hibler, 1976). These effects correspond closely to variations in Bunker's (1975) heat flux values and lead the oceanic response by a few years. The sense of the lag is consistent with the causal mechanism proposed by Worthington (op cit.), but in a sense, we are dealing with only a few major climatological events in the oceanic record (the minima in the late 1950's and mid 1970's and the maxima in the mid 1960's and late 1970's). Consequently all that can be claimed at this stage is that the sense and timing of the correlations are consistent with Worthington's hypothesis. The covariations are suggestive and bear further consideration, particularly in light of some mechanistic model and perhaps a larger data base.

9. Summary

The approach of obtaining annually averaged salinity on isopycnals for a 27 year record in the Sargasso Sea shows a smooth and significant modulation on decade timescales. The simple concept model that the high salinities are somehow driven by greater than normal heat loss (and hence watermass renewal) is supported by strong positive salinity-oxygen correlations in the upper thermocline. The lateral invasion of low salinity-high oxygen northern modes is seen in the IS-O_2 trend to negative correlation at $\sigma_\theta = 27.8$. A weaker, but still significant, anticorrelation between vertical stability and IS in the upper thermocline is again consistent with the model. Observations of ^3H-^3He systematics for two periods also indicate that IS is indeed an indicator of the degree of water mass renewal. Back calculation using the transient tracer-obtained oxygen utilization rate and the reconstructed (from IS-AOU) AOU record gives a pattern of variation in watermass renewal which, for the shallowest modes, have a range of a factor of two in ventilation rates. The relative amplitude of the watermass renewal variations decrease with increasing

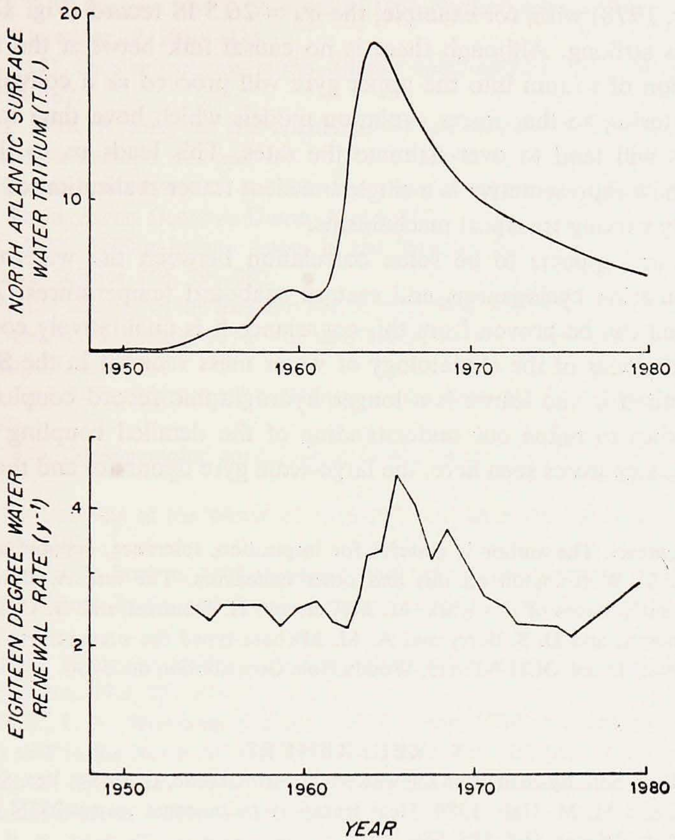

Figure 14. A comparison of the surface water tritium transient (after Dreisigacker and Roether, 1978) and Eighteen Degree Water (σ_θ = 26.5) renewal. The point is that the high degree of correlation, although not causal, will lead to biased results in transient tracer modelling of tritium if this kind of variation in watermass renewal is not accounted for in the model.

depth, and the relaxation timescales increase. This is consistent with increasing renewal timescales with depth.

The IS-depth correlations suggest that storage of denser and saltier water in the upper gyre may be the means of remembering or preconditioning the upper layers for more renewal, and the lagged covariance of IS with areal averaged heat fluxes lean toward lag times of 2-3 years.

The seasonal oxygen and stability records for the Eighteen Degree Water apparently show the imprint of the previous winter's renewal several months after the event. The time lags seen are consistent with the advective timescales from the formation regions to the monitoring stations.

What is important about the time series IS pattern is that if one compares the time history of North Atlantic surface water tritium concentrations (Dreisgacker and Roether, 1978) with, for example, the $\sigma_\theta = 26.5$ IS record (Fig. 14) the cross-correlation is striking. Although there is no causal link between the two records, the penetration of tritium into the upper gyre will proceed as a convolution of the two time histories, so that tracer evolution models which have time invariant ventilation rates will tend to over-estimate the rates. This leads us to the inevitable question of how representative is a single transient tracer realization of what appear to be strongly varying transport mechanisms.

Finally, there appears to be some correlation between the watermass renewal record and marine cyclogenesis and eastern seaboard temperatures. Although no causal relation can be proven from this covariance it is qualitatively consistent with Worthington's ideas of the climatology of water mass renewal in the Sargasso Sea. What is required in the future is a longer hydrographic record coupled with some synoptic studies to refine our understanding of the detailed coupling between the hydrographic signatures seen here, the large-scale gyre dynamics and the climate.

Acknowledgments. The author is grateful for inspiration, tolerance, support and encouragement from L. V. Worthington on this and other endeavors. The late A. Bunker was most helpful in the early stages of this work. M. McCartney, H. Stommel and G. C. Rooth all gave valuable comments, and D. S. Berry and A. M. Michael typed the manuscript. This work was supported by NSF Grant OCE 7-19815. Woods Hole Contribution no. 5050.

REFERENCES

Barrett, J. R. 1969. Salinity changes in the western North Atlantic. Deep-Sea Res. Supp., *16*, 7–16.

Bryden, H. L. and M. M. Hall. 1980. Heat transport by currents across 25°N latitude in the Atlantic Ocean. Science, *207*, 884–886.

Bunker, A. F. 1975. Energy exchange at the surface of the western North Atlantic Ocean. W.H.O.I. Report No. 75-3, 107 pp.

—— 1976. Computations of surface energy flux and annual air-sea interaction cycles of the North Atlantic Ocean. Mon. Wea. Rev., *104*, 1122–1140.

—— 1980. Trends of variables and energy fluxes over the Atlantic Ocean from 1948 to 1972. Mon. Wea. Rev., *108*, 720–732.

Bunker, A. F. and L. V. Worthington. 1976. Energy exchange charts of the North Atlantic Ocean. Bull. Amer. Met. Soc., *57*, 670–678.

Craig, H. 1969. Abyssal carbon and radiocarbon in the Pacific. J. Geophys. Res., *74*, 5491–5507.

Diaz, H. F. and R. G. Quayle. 1978. The 1976-1977 winter in the contiguous United States in comparison with past records. Mon. Wea. Rev., *106*, 1393–1421.

Dreisigacker, E. and W. Roether. 1978. Tritium and Sr-90 in North Atlantic surface water. Earth Planet. Sci. Lett., *38*, 301–312.

Ellis, J. S., T. H. Vander Maar, S. Levitus and A. H. Oort. 1978. The annual variation in the global heat balance of the earth. J. Geophys. Res., *83*, 1958–1962.

Fine, R. A. and H. G. Ostlund. 1980. Exchange times in the Pacific equatorial current system. Earth Planet. Sci. Lett., *49*, 447–452.

Frankignoul, C. 1981. Low frequency temperature fluctuation, off Bermuda. J. Geophys. Res., *86*, 6522–6528.

Gargett, A. E., T. B. Sanborn and T. R. Osborn. 1979. Surface mixing layers in the Sargasso Sea. J. Phys. Oceanogr., *9*, 1090–1111.

Hayden, B. P. 1981. Secular variation in the Atlantic coast extratropical cyclones. Mon. Wea. Rev., *109*, 159–167.

Iselin, C. O'D. 1936. A study of the circulation of the western North Atlantic. P. Phys. Oceanogr. and Meteorol., *4*, 101 pp.

—— 1939. The influence of vertical and lateral turbulence on the characteristics of waters at mid-depths. Trans. Amer. Geophys. Union, *3*, 414–417.

Jenkins, W. J. 1977. Tritium-helium dating in the Sargasso Sea: a measurement of oxygen utilization rates. Science, *196*, 291–292.

—— 1980. Tritium and ^{3}He in the Sargasso Sea. J. Mar. Res., *38*, 533–569.

Jenkins, W. J. and W. B. Clarke. 1976. The distribution of ^{3}He in the western Atlantic Ocean. Deep-Sea Res., *23*, 481–494.

Jenkins, W. J. and P. B. Rhines. 1980. Tritium in the deep North Atlantic Ocean. Nature, *286*, 877–880.

Lazier, J. R. 1980. Oceanographic conditions at O.W.S. Bravo, 1964-1974. Dyn. Atm. Ocean, *18*, 227–238.

Leetmaa, A. 1977. Effects of the winter of 1976-1977 on the northwestern Sargasso Sea. Science, *148*, 188–189.

Leetmaa, A. and A. F. Bunker. 1978. Updated draft of the mean annual windstress convergences in the Ekman layers and Sverdrup transports in the North Atlantic. J. Mar. Res., *36*, 311–322.

Mantyla, A. 1980. Electrical conductivity comparisons of Standard Seawater batches P29 to P84. Deep-Sea Res., *27A*, 837–846.

McCartney, M. S., L. V. Worthington and M. E. Raymer. 1980. Anomalous water mass distributions at 55W in the North Atlantic in 1977. J. Mar. Res., *38*, 147–172.

Mock, S. J. and W. D. Hibler. 1976. The 20-year oscillation in eastern North American temperature records. Nature, *261*, 484–486.

Montgomery, R. B. 1938. Circulation in the upper layers of southern North Atlantic deduced with use of isentropic analysis. P. Phys. Oceanogr. and Meteor., *65*, 55 pp.

Munk, W. H. 1966. Abyssal recipes. Deep-Sea Res., *13*, 707–730.

Ostlund, H. G., H. G. Dorsey and C. G. Rooth. 1974. GEOSECS North Atlantic radiocarbon and tritium results. Earth Planet. Sci. Lett., *23*, 69–86.

Pocklington, R. 1972. Secular changes in the ocean off Bermuda. J. Geophys. Res., *77*, 6604–6607.

Riley, G. A. 1951. Oxygen, phosphate and nitrate in the Atlantic Ocean. Bull. Bingham Oceanogr. Coll., *13*, 1–126.

Rooth, C. G. and H. G. Ostlund. 1972. Penetration of tritium into the North Atlantic thermocline. Deep-Sea Res., *19*, 481–492.

Schmitt, R. 1981. Form of the temperature-salinity relationship in the Central Water: evidence for double diffusive mixing. J. Phys. Oceanogr., *11*, 1015–1026.

Schroeder, E., H. Stommel, D. Menzel and W. Sutcliffe. 1959. Climatic stability of Eighteen Degree Water at Bermuda. J. Geophys. Res., *64*, 363–366.

Stommel, H. 1979. Determination of water mass properties of water pumped down from the Ekman layer to the geostrophic flow below. Proc. Natl. Acad. Soc. USA, *76*, 3051–3055.

Streeter, S. S. and N. J. Shackleton. 1979. Paleocirculation of the deep North Atlantic: 150,000-year record of benthic foraminifera and oxygen-18. Science, *203*, 168–171.

Stuiver, M. and H. G. Ostlund. 1980. GEOSECS Atlantic radiocarbon. Radiocarbon, *22*, 1–24.

Talley, L. D. and M. E. Raymer. 1982. Eighteen Degree Water variability. J. Mar. Res., *40* (Supp.), 757–775.

Warren, B. A. 1972. Insensitivity of subtropical mode water characteristics to meteorological fluctuations. Deep-Sea Res., *19*, 1–19.

Weiss, W. M., W. Roether and E. Dreisigacker. 1979. Tritium in the North Atlantic Ocean. "Behavior of tritium in the environment." IAEA-SM-232/98, 315–336.

Worthington, L. V. 1959. The 18° water in the Sargasso Sea. Deep-Sea Res., *5*, 297–305.

—— 1962. Evidence for a two gyre circulation in the North Atlantic. Deep-Sea Res., *9*, 51–67.

—— 1968. Genesis and evolution of water masses. Met Monogr., *8*, 63–67.

—— 1971. Anticyclogenesis in the oceans as a result of outbreaks of continental polar air, *in* Studies in Physical Oceanography—A Tribute to George Wust on his 80th birthday, A. L. Gordon, ed., Gordon and Breach, N.Y., 194 pp.

—— 1972. Negative oceanic heatflux as a cause of watermass formation. J. Phys. Oceanogr., *2*, 205–211.

—— 1976. On the North Atlantic circulation. The Johns Hopkins Oceanogr. Studies No. 6, Baltimore, Johns Hopkins University Press, 110 pp.

—— 1977. Intensification of the Gulf Stream after the winter of 1976-1977. Nature, *270*, 415–417.

Wunsch, C. 1972. The spectrum from two years to two minutes of temperature fluctuations in the main thermocline at Bermuda. Deep-Sea Res., *19*, 577–593.

Received: *17 August, 1981*; revised: *8 February, 1982.*