JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 107, NO. C3, 3021, 10.1029/2000JC000550, 2002

Stability of the Mediterranean's thermohaline circulation under modified surface evaporative fluxes

Paul G. Myers[1]

Department of Physics and Physical Oceanography, Memorial University of Newfoundland, St. John's, Newfoundland Canada

Keith Haines

Environmental Systems Science Centre (ESSC), Reading University, Reading, UK

Received 13 July 2000; revised 27 April 2001; accepted 7 August 2001; published 23 March 2002.

[1] A series of experiments with an ocean general circulation model of the Mediterranean forced by artificial (but realistic) surface fluxes of heat and fresh water is performed over 100 year periods. The model has a stable thermohaline circulation under the baseline fluxes. Small increases/decreases in excess evaporation (±8%) produce a linear strengthening/weakening of this thermohaline circulation with more/less water formation and strait transports. The water properties do not change much, and the responce is consistent with submaximal exchange at Gibraltar, where changes in volume and freshwater transport can be achieved mainly by changing the interface level between inflowing and outflowing waters. Larger changes in excess evaporation lead to nonlinear responses for increasing and decreasing values that are quite different. Reducing evaporation (−20%) causes both western and eastern deep water formation to cease, leaving a shallower intermediate circulation only. Increasing evaporation (+25%) initially strengthens the overturning but also produces a hydraulic jump-like feature at Gibraltar, which mixes inflowing and outflowing waters, decreasing the salt flux out of the basin. As a result, the basin salinity increases greatly over 140 years, until the overturning collapses and the Gibraltar mixing ceases, leaving a shallow intermediate circulation above highly saline and stagnant deep waters. The possible relevance of these experiments to recent changes in Mediterranean fluxes and water properties is discussed. *INDEX TERMS:* 4243 Oceanography: General: Marginal and semienclosed seas; 4255 Oceanography: General: Numerical modeling; 4215 Oceanography: General: Climate and interannual variability (3309); *KEYWORDS:* Mediterranean, Strait of Gibraltar, multiple equilibria, hydraulic control, flux forcing

1. Introduction

[2] The Mediterranean is a small marginal sea located between Europe and Africa. Despite its marginal nature it is an important basin because it possesses an active thermohaline circulation, with deep convection and water mass formation, and therefore it has considerable similarity to the global ocean. With its small size, however, its overturning timescale is very amenable to numerical simulation, making it an excellent test bed for ideas on the large-scale oceanic circulation.

[3] The Mediterranean is a concentration basin in which evaporation E exceeds precipitation P and river runoff R by about 70 cm yr^{-1}, [*Gilman and Garrett*, 1994; *Macdonald et al.*, 1994; *Myers and Haines*, 2000] while undergoing a net surface heat loss of ~6 W m^{-2} [*Macdonald et al.*, 1994; *Bethoux*, 1979]. Warmer, fresher water enters at the surface from the Atlantic through Gibraltar, and colder saline water leaves below. This two-layer exchange system at Gibraltar has been much discussed since it is probably subject to hydraulic control [e.g., *Bryden and Kinder*, 1991]. The exchange system at Gibraltar has considerable variability on tidal timescales [e.g., *Candela*, 1991; *Wesson and Gregg*, 1994] and perhaps on longer timescales such as seasonal and interannual, although these have not been well observed.

[4] The water formation and mixing processes in the basin are known to show considerable variability. Water formation takes place in winter, and several authors have noted considerable variability in winter convection depths and properties around the basin [e.g., *Leaman and Schott*, 1991; *Sur et al.*, 1993]. The interannual variability in air-sea fluxes responsible are more difficult than water properties to observe accurately, but *Garrett et al.* [1993], and more recently S. A. Josey (personal communication, 2000), have calculated interannual variability in $E - P$ of nearly 10% (and in heat flux of ±15 W m^{-2}). Relating water exchange at Gibraltar with the buoyancy budget of the basin is a simple procedure if the so-called steady "overmixing" assumption is made, in which the density of the outflowing water is the minimum required to drive sufficient net exchange to balance the basin buoyancy loss. *Bryden and Kinder* [1991] have shown that the Mediterranean may be close to this state. However, if the basin is not "overmixed," the relationship between basin buoyancy loss and the exchange system at Gibraltar has more freedom. It is particularly important to explore this freedom in the context of changing conditions over the Mediterranean in recent decades.

[5] *Bethoux et al.* [1990], *Leaman and Schott* [1991], and *Rohling and Bryden* [1992] have noted that both deep water potential temperatures and salinities have increased in the Gulf of Lions this century, with the rate of increase accelerating after 1960. *Leaman and Schott* [1991] suggested that the changes were due to changes in the Levantine Intermediate Water (LIW) from the

[1] Now at Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta, Canada.

0148-0227/02/2000JC000550

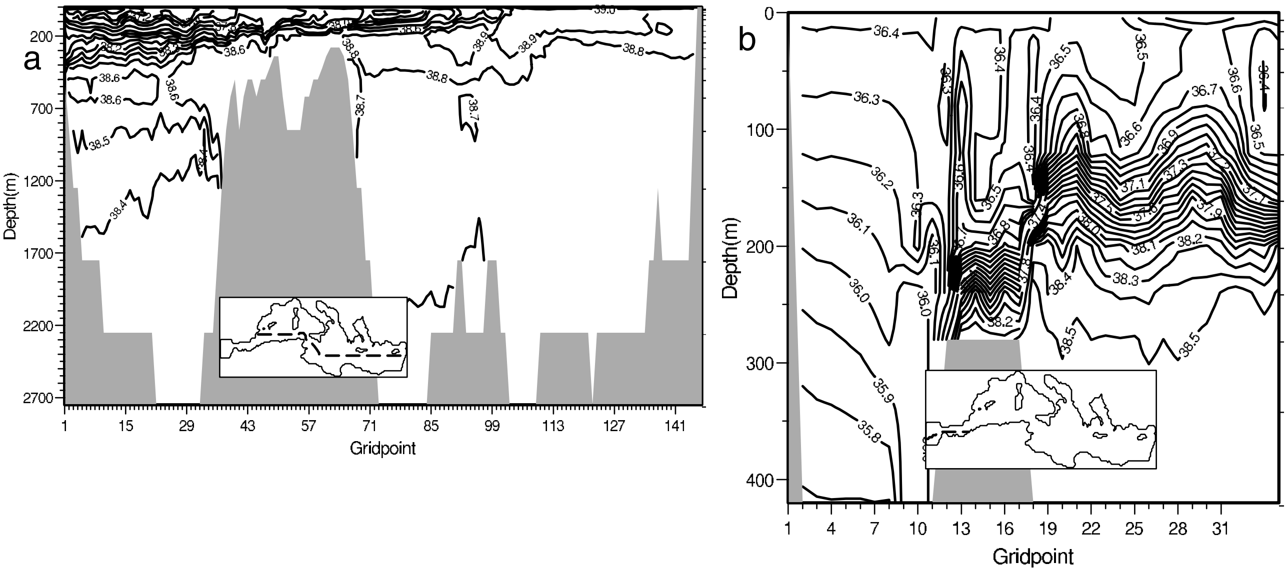

Figure 1. Annually averaged salinity cross sections through (a) the entire basin and (b) the Strait of Gibraltar after 100 years of integration for our baseline experiment. The contour interval is 0.1 psu.

eastern basin, a view also supported by *Rohling and Bryden* [1992]. *Bethoux and Gentili* [1996] discussed the long-term changes in buoyancy forcing over the basin, particularly those due to changing river inputs, e.g., the damming of the Nile. Recently, in the eastern basin, very large hydrological changes were observed between 1987 and 1995 [e.g., *Roether et al.*, 1996], much larger than anything observed in the west.

[6] The role of changing buoyancy fluxes in causing the alterations in the Mediterranean thermohaline circulation and water properties is still the subject of much debate. In this paper we take advantage of recent modeling studies of the thermohaline circulation to determine the sensitivity of an ocean general circulation model (OGCM) to changes in buoyancy forcing. *Myers and Haines* [2000] showed that it is possible to run a 1/4° 19-level OGCM of the Mediterranean for centuries using only flux forcing at the surface (seasonally varying) and to retain a reasonably realistic thermohaline circulation with all eastern and western water masses represented. The fluxes used were diagnosed from a previous model run in which the surface properties were relaxed to observations, and these fluxes are realistic on the basin scale. Here we investigate the effect of modifying these fluxes, in particular, the $E - P - R$, on the OGCM circulation. We are interested in exporing the relation between water formation rates, mixing, and Gibraltar exchanges and interpreting the results, where possible, in terms of simple hydraulic and budget models of the Mediterranean. We are also interested in how large a change in buoyancy forcing is required to change the qualitative nature of the thermohaline circulation. This leads to some surprising results. The model and its spin-up are described in section 2. The cases with increased and decreased net evaporation and heat fluxes are described in section 3. An interpretation of linear thermohaline stability is given in section 4, while nonlinear responses are examined in section 5. We conclude with a discussion in section 6, including an examination of the relevance of this work to the present day Mediterranean.

2. Model and Spin-Up

[7] The model used is the Modular Ocean Model—Array (MOMA), a Bryan-Cox-Semtner-type OGCM, using the *Killworth et al.* [1991] free-surface scheme. The tracer advection schemes are modified to include the *Gent and McWilliams* [1990] (hereinafter referred to as GM) eddy parameterization and a flux -limiter by [*Stratford*, 2000], based on the work of *Thuburn* [1996]. The basic model is described in greater detail by *Webb* [1993] and has been used in the Mediterranean to study changes in the basin's thermohaline circulation, both for the present [*Myers et al.*, 1998a; *Myers and Haines*, 2000] and for past climates [*Myers et al.*, 1998b; *Myers and Rohling*, 2000].

[8] The model setup is based on the work of *Haines and Wu* [1995] and *Wu and Haines* [1996, 1998]. Basin resolution is 0.25° × 0.25° with 19 vertical levels, concentrated in the upper water column to resolve the thermocline. The horizontal biharmonic viscosity is $A_h = 1.5 \times 10^{18}$ cm^4 s^{-1}; the vertical momentum diffusion is $A_v = 1.5$ cm^2 s^{-1}; the GM thickness diffusion parameter is 2.0×10^5 cm^2 s^{-1} and the maximum (reciprocal) slope of isopycnals is 100.0. Convective adjustment uses the complete convection scheme of *Rahmstrof* [1993]. To handle the exchanges with the Atlantic, a small box is added outside of Gibraltar, where the temperature and salinity are relaxed on a 1 day timescale to the climatological values of *Levitus* [1982]. Wind stress data are based upon a monthly 7 year climatology (1986–1992) from the European Centre for Medium-Range Weather Forcasting.

[9] To produce our initial baseline Mediterranean experiment, the model was spun up from rest using relaxation surface T, S conditions. Both the initial hydrography and monthly varying surface relaxation conditions were derived from the Mediterranean Oceanic Data Base MED5 [*Brasseur et al.*, 1996]. The model was integrated for 100 years until a steady state was reached, and monthly surface fluxes were diagnosed over the last 15 years of integration. Following this, all relaxation was removed, and the model was forced only with the diagnosed monthly surface fluxes for another 100 years. The model remained stable under the fluxes, although variability was enhanced [*Myers and Haines*, 2000].

[10] Figure 1a is a transect across the basin from the baseline after 100 years showing that the model does a good job of reproducing and maintaining an LIW layer in both basins. In the east the model produces LIW with a core density of $\sigma_\theta = 29.05$ and a salinity of 39.0 psu [*Myers and Haines*, 2000]. This LIW preconditions eastern deep water formation in the Adriatic and also passes westwards at Sicily. As LIW passes through Sicily, it mixes and sinks into the Tyrrhenian, with a salinity and depth

Table 1. Annual Surface Heat and Freshwater Fluxes Diagnosed From the Final 15 Years of an Initial Model Spin-Up Under Restoring Boundary Conditions[a]

	Heat Flux, W m^{-2}	$E - P - R$, cm yr^{-1}
MED	−5.8	76
WMED	−7.1	94
EMED	−5.1	66
ADR	−41	−20

[a] The regional abbreviations are MED, the basin average; WMED, the western Mediterranean; EMED, the eastern Mediterranean (including the Adriatic and Aegean); and ADR, the Adriatic.

range broadly consistent with observations [*Sparnocchia et al.*, 1999]. In the western basin much of the LIW circulates cyclonically, reaching the Gulf of Lions, where western deep water is produced. Some LIW is also mixed westward south of Sardinia directly toward Gibraltar. Comparison with *Send et al.* [1999, Figure 1] shows that although the model LIW is slightly too salty and deep, a reasonable LIW structure is maintained near the Strait of Gibraltar (Figure 1b). The end of this experiment was used as starting point for the flux modification experiments described below.

[11] The annual average fluxes diagnosed from the model are shown in Table 1 and compare very well with observational estimates on the basin and subbasin scale. Monthly flux values from this run are also shown by *Myers and Haines* [2000]. The seasonal heat fluxes are in phase with those from the SOC climatology [*Josey et al.*, 1998] but about half the amplitude, although this does not greatly affect the water formation. The weaker cycle is due to lack of a surface mixing scheme and the use of GM mixing, which means that in summer the seasonal thermocline is very shallow. In winter, then, less heat needs to be removed to break through the summer thermocline. In all experiments here these fluxes are modified in a very simple manner.

3. Sensitivity to Net Evaporation

3.1. Multiplicative Changes

[12] In these experiments, for each grid point and each month the net evaporation is multiplied by a factor to produce the change desired. This has the effect of increasing (or decreasing) spatial and temporal variability in net evaporation by the same factor. For example, to increase excess evaporation by 5%, the original fluxes are all multiplied by 1.05. This form of flux change will capture large-scale atmospheric changes but will necessarily miss local changes, which may have significant effect on regional water formation. However, paleoclimate model experiments with regional flux changes show that although local circulation and hydrography depend on local fluxes, provided the circulation does not qualitatively change, it is only the net buoyancy that affects the Gibraltar exchange [*Myers and Rohling*, 2000; *Myers*, 2002]. The experiments performed are summarized in Table 2.

[13] Consider an increase in the excess evaporation of 8%. Starting from the end of the run with the original fluxes, the model is integrated for a further 100 years with the enhanced fluxes. There is little change to the thermohaline circulation of the basin, with all the major water masses still being formed. There has been an increase in the strength of the thermohaline overturning cell, mainly in the deep water cell of the eastern basin. This can be seen in the overturning stream function (Figures 2a and 2b) and is reflected in the increased Gibraltar volume and freshwater transports in Table 2 (The freshwater transport is inferred from the salt transport since the model is volume conserving. In a steady state this will balance the net $E - P$, but any imbalance implies a changing salt content of the basin and not a change in freshwater volume.) The outflow salinity, measured at the eastern end of the strait, hardly changes. Although the value appears to be very low, Figure 1b shows that the model flow has a wide transition zone so considerable mixing has already occurred at the strait entrance. The values reported for the outflow salinity are the average of all grid boxes with a negative (outgoing) velocity component and thus include a fair volume of inflowing Atlantic water, significantly reducing its salinity. The deep water salinity just inside the basin reaches 38.5 psu, which is well in accord with observed values.

[14] The freshwater transport at Gibraltar increases by very close to 8% with an 8% increase in the excess evaporation, showing that the freshwater/salt budget of the basin is rebalanced after 100 years. The volume transport at Gibraltar also increases by very close to 8%, with the outflow salinity only marginally increasing. These observations have implications for how the basin as a whole responds to changing buoyancy fluxes. Water formation in the basin can respond in two ways to an increased net buoyancy loss. The basin can form more water of the same density or form the same amount of water with a higher density, or somewhere in

Table 2. Summary of the Experiments Performed, Showing the Changes in Surface Net Evaporation, Basin Average Salinity, Outflow Salinity at Gibraltar, Transports at Gibraltar, and Changes in Basin Overturning Strengths[a]

Experiment	Year	Basin Average S, psu	Gibraltar Volume Transport, Sv	Gibraltar Freshwater Transport, Sv	Outflow S, psu	Zonal Overtuning Strength, Sv	Eastern Deep Water Formation, Sv	Western Deep Water Formation, Sv
Baseline	100	38.53	1.59	0.058	37.47	1.45	0.30	0.20
+5%	100	38.57	1.67	0.061	37.49	1.50	0.35	0.20
+8%	100	38.59	1.71	0.063	37.50	1.55	0.50	0.20
+25%	120	39.25	1.83	0.047	37.11	1.75	0.70	0.40
+25%	200	39.15	1.54	0.073	37.69	1.50	0.00	0.00
−5%	100	38.51	1.52	0.054	37.38	1.35	0.10	0.10
−8%	100	38.48	1.45	0.053	37.46	1.30	0.10	0.10
−20%	100	38.43	1.48	0.048	37.36	1.30	0.00	0.00
+3.5 cm yr^{-1}	100	38.58	1.51	0.061	37.50	1.50	0.35	0.20
H+5%	100	38.54	1.57	0.058	37.47	1.45	0.40	0.20
H−5%	100	38.54	1.53	0.055	37.48	1.35	0.10	0.10

[a] The year indicates the length of each integration with values calculated over the last 10 years. Experiments prefixed with an H are modified heat flux experiments.

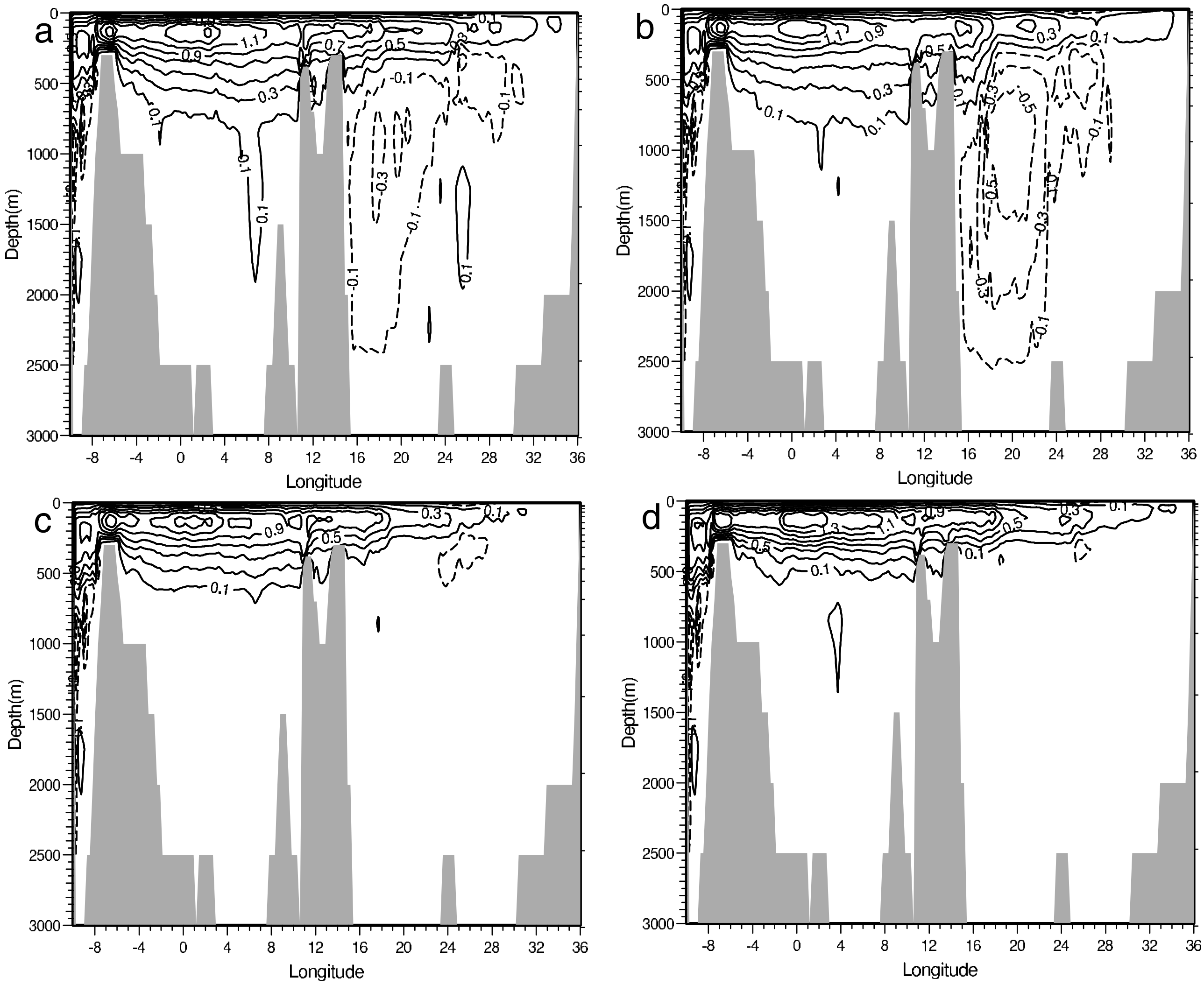

Figure 2. Annually averaged zonal overturning stream function from the final year of integration of (a) baseline, (b) +8%, (c) −20%, and (d) +25%. The contour interval is 0.2 Sv.

between. The production of more water in this model corresponds to a speed up of the thermohaline circulation in response to the extra forcing.

[15] This interpretation is confirmed by the water formation rates when measuring the change in the volume of water with a salinity greater than some threshold from the beginning to end of each winter and by the increasing strength of the zonal overturning circulation. The correspondance between net evaporation, water formation, and Gibraltar transports holds well for changes up to ±8% in net evaporation, with results for ±5 and ±8% being shown in Table 2. For flux changes beyond this range the basin circulation becomes qualitatively different, and we refer to this behavior as nonlinear. Two experiments are included, with evaporation changes of +25 and −20%. In section 4 we discuss the interpretation of the linear regime in terms of the classical two-layer description of the basin, and in section 5 we consider the nonlinear regimes.

3.2. Homogeneous Changes

[16] One experiment in which the evaporation was increased uniformly across the basin in all seasons was also performed. A 5% evaporation increase corresponds to a fixed 3.5 cm yr^{-1} of extra evaporation applied at all grid points spread over all seasons. The general behavior in this experiment is similar to that from the +5% multiplicative experiment. The basin freshwater budget is balanced, and the mean salinity of the basin is virtually identical to the other run. A difference plot of the salinity from these two experiments (not shown) has only localized differences, in the eastern basin deep water, because of the higher salinities this flux change implies for the Adriatic. The large-scale circulation and hydrography are also the same. The main differences are in the Adriatic and eastern basin where the multiplicative increase experiment enhances the present precipitative nature of the Adriatic, but the homogeneous change experiment produces a basin that has less fresh water near the surface. This leads to an increase in convection in the basin and thus more production of Eastern Mediterranean Deep Water in the homogeneous change experiment. There was also a small drop in the Gibraltar transport, for reasons that we do not yet understand.

3.3. Sensitivity to Net Heat Flux

[17] In these experiments the basin-averaged heat flux was increased and decreased by 5%, in the multiplicative sense. This also amplifies or damps the seasonal cycle, which is very large for heat fluxes as shown in Table 1.

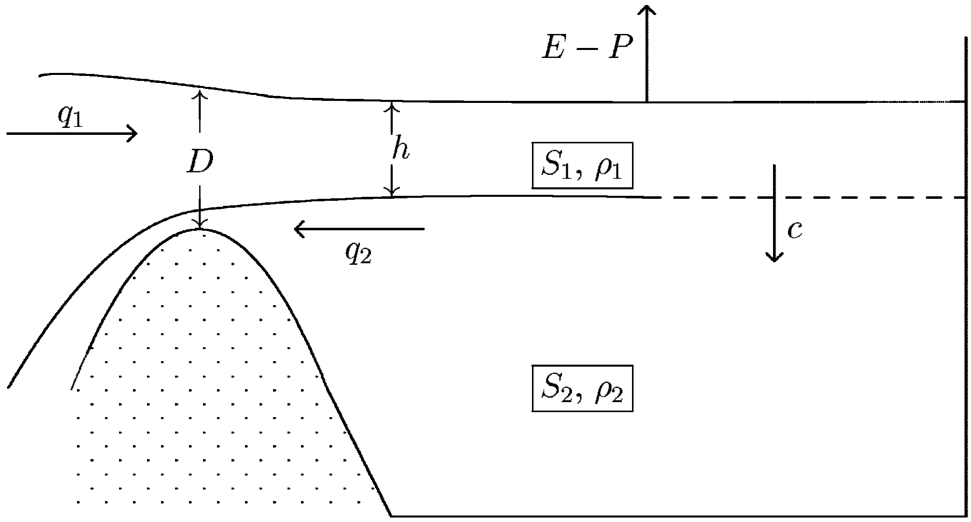

Figure 3. Schematic showing water formation in a two-layer basin. The transports q_1 and q_2, net evaporation $E - P$, and water formation rate c are in sverdrups. The slope on the top surface, down toward the basin, is greatly exaggerated and is, in reality, only a few centimeters.

[18] The increase or decrease in the basin average heat loss has little effect on the salinities, as seen in Table 2. However, changing the net heat loss from the basin does have a significant effect on the deep water formation, both in the Adriatic and the Gulf of Lions. Increasing heat loss increases the water formation in the Adriatic, while decreasing the heat loss decreases the water formation in both the Adriatic and Gulf of Lions substantially. However, there is little change to intermediate water formation and thus to the hydraulically controlled flow at Gibraltar. The reason for this is that although salinity is the main driving force for the large-scale thermohaline circulation, it is the wintertime cooling that actually drives the localized deep water formation. Thus changes in the heat fluxes have a smaller effect on the main zonal thermohaline cell, consistent with their contribution to the net buoyancy budget, but large effects on the deep water cells within each subbasin.

4. Interpretation of Linear Thermohaline Sensitivity

4.1. Exchanges at Gibraltar

[19] If we return to the basin response to small changes in $E - P$, we can interpret the OGCM results in terms of a two-layer hydraulically controlled Mediterranean model. Figure 3 illustrates the situation where q_1, q_2, and c are the inflow rate, outflow rate, and water formation rate, respectively. For convenience we assume that heat fluxes are not involved in the basin budgets. Provided the basin is not in a maximum exchange state it is possible to increase or decrease $E - P$, c, q_1, and q_2 all by say 8% without changing S_1, S_2. The interface level above the sill, $D - h$, will then change to alter the transports q_1, q_2 by the requisite amount.

[20] We examine this by looking at density sections through the Strait of Gibraltar in Figure 4 for the baseline and 8% increase experiments. It is clear that despite the crudeness of the OGCM resolution the two-layer exchange is not a bad approximation to the exchange at Gibraltar, although the transition zone is wide. From the density difference in Figure 4c one can see that there has been a rise in the interface position in the Alboran between the two experiments with no significant increase in the deep water densities. Note thatthis type of response is only possible in the submaximal strait exchange regime where changes in interface position can change the transport. Even in the submaximal regime, the model could have responded differently to the increased fluxes by increasing the density of the new water rather than the volume. The fact that it does not is extremely important. By converting much of the increased net E into a greater water volume the system moves rapidly toward a maximum exhange situation. In section 4.2 we show that this explains why the net E cannot be increased by more than about 8%, as this appears to lead to an analogue of the maximal exchange situation at Gibraltar.

4.2. Water Formation Within the Basin

[21] Returning to the water formation sites within the basin, the impact of the changed fluxes is more complicated. With an 8% increase in the excess evaporation the basin continues to form LIW with the same core density in the eastern basin of $\sigma_\theta \sim 29.05$. However, the area of water formation where this isopycnal outcrops in winter gets larger. The mean salinity of this intermediate water also increases from 38.98 to 39.02 psu, while the mean temperature increases to compensate from 15.6° to 15.8°C. The increase in intermediate water temperature is a sign of the speeding up of the water formation rate given that the heat fluxes are the same. These changes persist downstream in the LIW core in the Ionian and western basins. This is consistent with the changes to LIW formation found by *Lascaratos et al.* [1993] in a layer model of the Levantine. Here we show that increased salinity fluxes have an effect similar to cooling fluxes; it is the fact that there is a change in the new buoyancy forcing that is significant.

[22] The reason why extra buoyancy loss leads to more water formation of the same density is due to the way intermediate waters are formed. The areas of intermediate water formation are not geographically constrained and can therefore increase in response to extra forcing. Convection can also occur to greater depths before dispersal within baroclinic eddies, leading to a larger volume of water formed. The behavior is different from that at the deep water formation sites in the Adriatic and Gulf of Lions. These areas are highly geographically confined, and the convection often reaches to the bottom, so there is little scope for changing the volume of waters formed. The 8% increase in evaporative fluxes increases the model Eastern Mediterranean Deep Water densities from 29.23 to 29.28 and those of the Western Mediterranean Deep Water (WMDW) to 29.14 from 29.12. Table 2 shows that the mean salinity and density of the whole Mediterranean has increased slightly, and since the outflow density has not changed, we interpret this as due to an increase in deep water density and an increased volume of deep and intermediate waters, i.e., a raised thermocline, as seen in Figure 4c. It is the intermediate water that

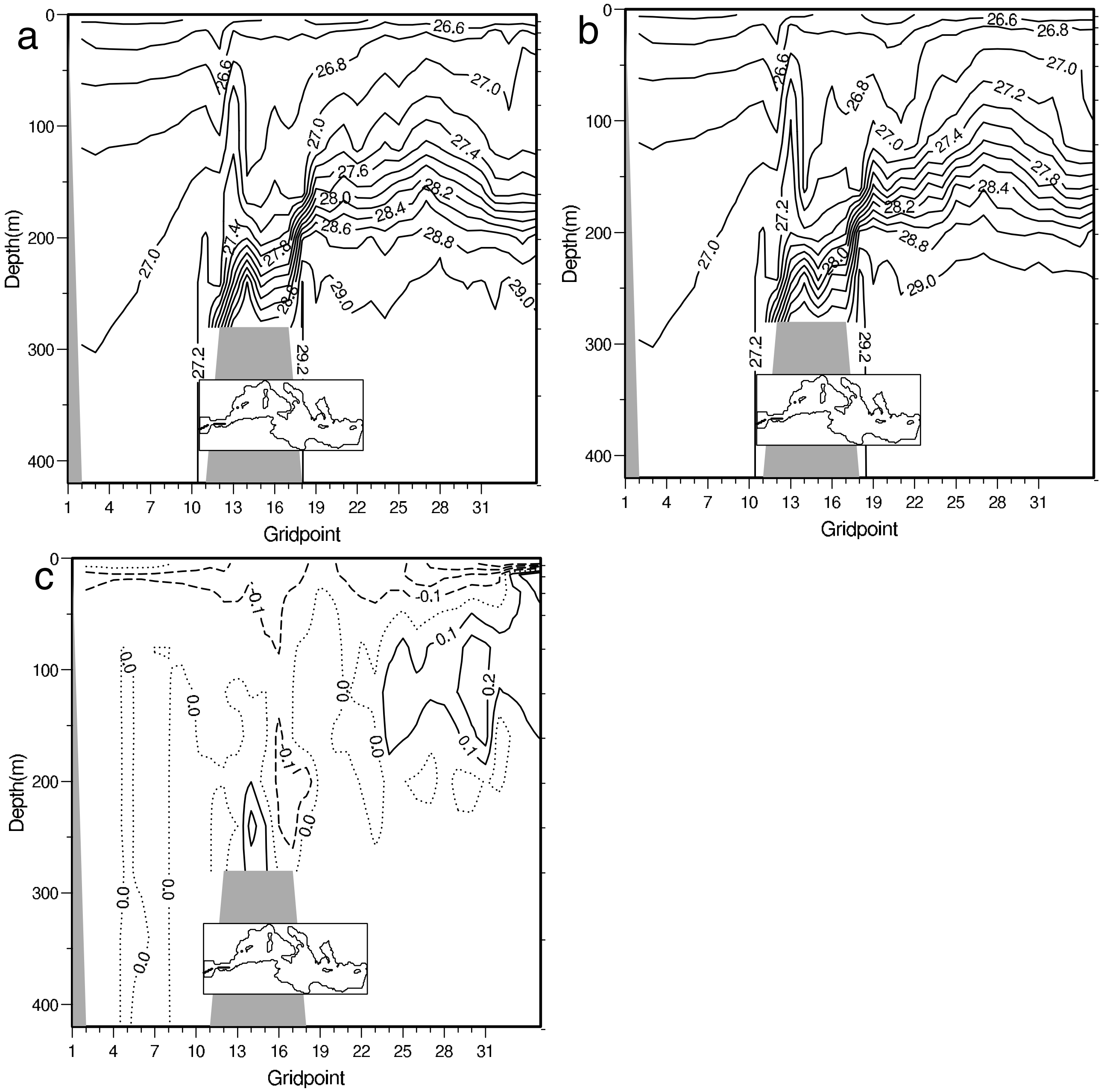

Figure 4. Annually averaged density cross sections through the Strait of Gibraltar after 100 years of integration for (a) baseline, (b) +8% experiment, and (c) a different plot of Figure 4b–Figure 4a. The contour interval is 0.2 kg m^{-3} in Figures 4a and 4b and 0.1 kg m^{-3}.

makes up the bulk of the outflow, and therefore density changes are not seen at Gibraltar.

5. Interpretation of Nonlinear Thermohaline Responses

5.1. Decreasing Buoyancy Flux

[23] When the net evaporation is decreased by 20%, convection decreases greatly in intensity. The LIW formation ceases because of the decrease in salinity of the Modified Atlantic Water entering the Levantine. Only limited convection occurs in the Aegean, and without the LIW, wintertime cooling in the northern Ionian leads to formation of a cold, fresh, intermediate water mass. Adriatic deep convection still occurs, but without the preconditioning from LIW the water is no longer dense enough to sink to the bottom after overflowing the sill at Otranto, and instead, it occupies an intermediate layer between 1000 and 1500 m. The loss of salt from the LIW also weakens convection in the Gulf of Lions, limiting it to a depth of 1200 m. Although there is no convection to the bottom occurring, the stratification in the deep waters of both basins remains very weak.

[24] Table 2 shows that despite the 20% evaporation decrease, the Gibraltar transport has slightly increased from the value it had with an 8% decrease. The reason for this can be seen in the zonal overturning stream function (Figure 2c). With deep water formation stopped in both eastern and western basins the thermohaline circulation now consists only of a shallow intermediate cell. The strength of this cell is virtually unchanged, but the shallower depth is very evident from the figures. This shallower cell remains stable

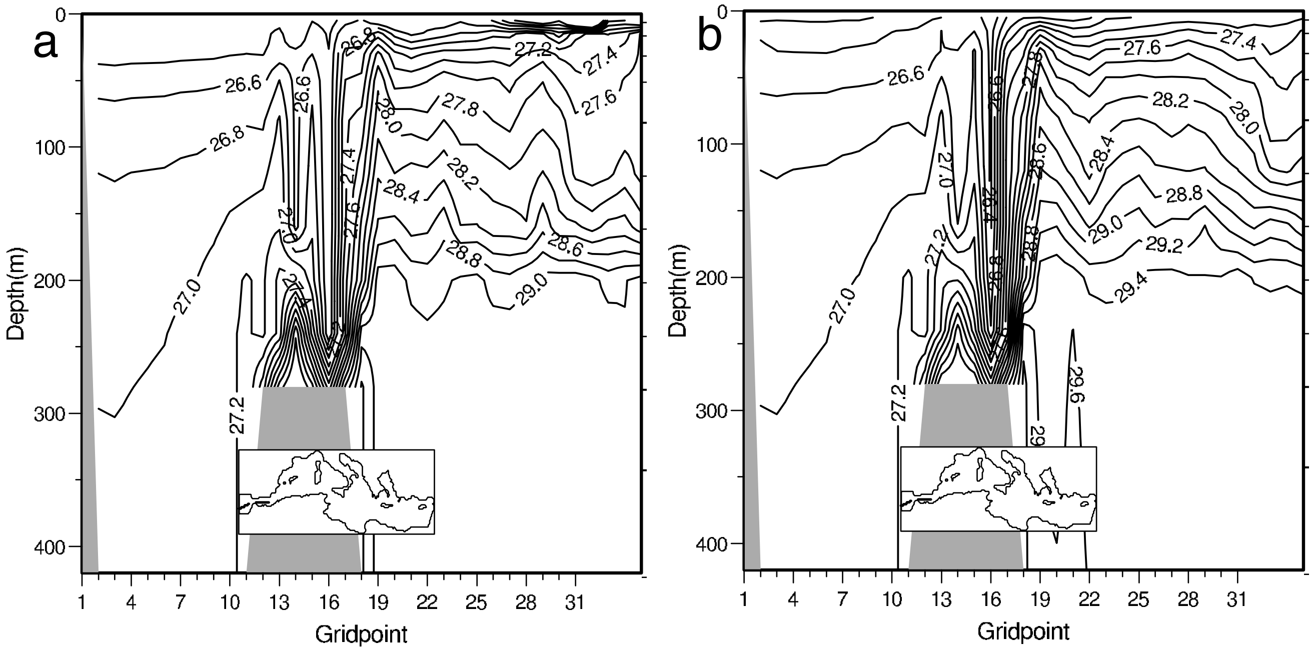

Figure 5. Annually averaged density cross sections through the Strait of Gibraltar with an increased evaporation of 25% (a) after 20 years and (b) after 100 years of the experiment. The contour interval is 0.2 kg m^{-3}.

and is consistent with the box modeling studies of G. Korres and A. Lascaratos (personal communication, 1999). Despite the very different mixing regime inside the basin the water exiting at Gibraltar has similar properties (albeit fresher as the outflowing LIW is now replaced with other eastern intermediate waters that undergo similar mixing processes along their path to Gibraltar), and there is little sign of the loss of deep water formation in the measured strait quantities.

5.2. Increasing Buoyancy Flux

[25] In section 4 we showed that the linear response of the basin to increasing net evaporation was driving up the interface position inside the strait, which can lead to a maximal exchange situation. Maximal exchange is the largest transport that can be driven through the straits for a given density difference between the inflowing and outflowing waters. Once maximal exchange is reached, further changes in interface height cannot change the strait transports; therefore the basin must respond by increasing the salinity and density of the outflow.

[26] It is possible to calculate the maximum exchange transport in the idealized situation of nonrotating frictionless two-layer exchange with the densities observed in Figure 4 and the strait cross-section from the model. With a 25 km wide and 280 m deep rectangular strait with exchange densities of 27 and 29.1 kg m^{-3} the maximum exchange is 2.4 Sv, well above the values observed in Table 2. It is unlikely that the OGCM could reproduce such a strong flow, but we can look for other signs of the "effective" maximum exchange state.

[27] We expect the onset of the effective maximal exchange to be signaled by the development of a hydraulic jump inside the Mediterranean basin just inside the straits. Remarkably, we do detect a feature that could be a hydraulic jump at around the time that the basin response becomes nonlinear. When a 25% increase in net evaporation is introduced, the basin average salinity and convective activity initially increases, but the behavior does not remain linear. Figure 5a shows an instantaneous section 20 years into this run, and Figure 5b shows the annual average section after 100 years. Figures 5a and 5b show almost vertical density contours just inside the strait with clearly a great deal of mixing going on. This feature always appears within a few decades when the net evaporation is increased by more than about 10%. An additional experiment (not shown) with a 10% increase shows this feature after an ~50 year time delay; otherwise, the behavior is similar to the 25% experiment in Table 2. Smaller increases in net evaporation mean a slower interface rise toward the effective maximal state, but once the hydraulic jump and associate mixing are initiated, the subsequent evolution is highly nonlinear. Given that the main response of the submaximal basin to an evaporation increase is to produce a larger volume of dense water, it is reasonable to conclude that the nonlinear regime is produced by the onset of a hydraulic jump at the effective maximal exchange point.

[28] While identifying the feature in Figure 5 as a hydraulic jump, we must state that we have no reason to believe that a real hydraulic jump would behave in exactly the same way since the model resolution is far too poor to be convincing in its treatment of such processes. It is a remarkable feature, and we have found very few references to numerical hydraulic jump modeling in the literature with which to compare our results. For a rectangular strait the jump should appear when the interface level inside the basin is shallower than about 35% of the sill depth (about 110 m) [*Dalziel*, 1991]. A fuller investigation of this phenomenon is currently underway and will be presented separately. For now we note that the most important impact on the thermohaline circulation, which explains the results summarized in Table 2, is that it leads to strong mixing between the inflowing and outflowing water masses. However, the degree of mixing that should occur in such jumps is not well documented.

5.3. Increasing Buoyancy Flux: Basin Circulation

[29] The consequences for the basin of the 25% increase in net evaporation can be seen in the higher densities of the near-surface water properties inside the Mediterranean in Figure 5, when compared with Figure 4, as well as the high density of the deep water in the Alboran Sea in Figure 5b. Table 2 shows that after 120 years, while the Gibraltar volume transport has greatly increased, the freshwater/salt transport is still low (0.047 Sv) and is not balanced (an import equivalent of 0.072 Sv is needed). This

explains why the average salinity of the basin has increased greatly and is still increasing.

[30] As the salinity of the basin deep water slowly increases, the freshwater transport at Gibraltar also increases, despite the mixing. The salinification of the basin continues for about 140 years with deep water formation in both basins remaining operative. During the early part of the run the steady increase in salt within the eastern basin leads to progressively deeper winter mixed layers everywhere, and within 10 years there is a complete breakdown of stratification, with nearly the entire basin north of 34°N in the eastern basin and 40°N in the western basin, convecting each winter. This causes the significant increase in deep salinity as the deep waters of Adriatic origin are replaced. This process continues for a few years until the eastern deep waters are replaced, and a new circulation pattern develops by year 15–20 of the integration.

[31] Convection to the bottom now occurs only in the Rhodes gyre, within the main area of the basin, forming Levantine Deep Waters with properties around S = 39.22 and T = 14.1°C. This water mass fills the deep Levantine but does not immediately cross west of the Cretan rise. Deep convection also continues to occur in the Adriatic with a colder and fresher water mass overflowing Otranto to provide deep water to the Ionian basin. During years 20–90 the importance of the Adriatic source decreases as the salinity, and thus density, of the Levantine waters rise, but the Adriatic source does persist. Since deep convection occurs in the Levantine, an intermediate saline water mass is formed in the Cretan sea, and this is exported westward and northward, penetrating the Adriatic and helping to maintain convection there. This Cretan water thus plays a similar role to LIW in the baseline experiment. It is interesting to note that this sequence of events is similar to those inferred from observations by *Klein et al.* [2000] and by *Lascaratos et al.* [1999] in a recent modeling study following the recent eastern Mediterranean transient. They described decreases Adriatic ventilation in the early 1990s after the supply of LIW to the Adriatic was cut off and diverted into the Aegean Sea. The Adriatic convection restarted in the late 1990s after salty intermediate water from the Aegean started to increase Adriatic salinity again.

[32] By about year 130 the outflowing salinity has increased sufficiently for the basin to reach a salt balance. At this point the deep water formation collapses, and a shallow intermediate overturning circulation is established, shown at year 200 in Figure 2d, which is quite similar to the circulation in Figure 2c. However, the abyssal waters are much denser now, with a very strong pycnocline between the ventilated and unventilated regions of the basin, making it highly unlikely that ventilation will be reestablished.

6. Discussion

6.1. Relevance to Today's Circulation

[33] The most relevant results for the present-day Mediterranean relate to the model response to a net increase in $E-P$. The model begins in a submaximal state where the main response to small increases in $E-P$ is to increase the water formation and overturning rates of the basin. This leads to only small changes in the basin water properties at Gibraltar. However, a natural consequence of this is to raise the level of the deep waters inside the basin, which causes the Gibraltar flow to move toward maximal exchange. This linear dependence of the transport and the rate of water formation is consistent with the idea of *Hopkins* [1999] that the transport is driven by the internal pressure gradient through the strait.

[34] The approach to maximal exchange is accompanied by a large increase in mixing between inflowing and outflowing waters, which reduces the salinity of the outflowing waters. So, whereas in the submaximal regime the basin salinity hardly responds to small $E-P$ changes, when the maximal regime is approached, the basin salinity increases very rapidly. While the model cannot be trusted to handle the mixing at the straits quantitatively correctly, the result that mixing would rapidly increase in the maximal exchange state is a very realistic possibility.

[35] *Leaman and Schott* [1991] and *Rohling and Bryden* [1992] showed that small increases in deep water salinities have occurred on decadal timescales, with changes in the WMDW of the order of 0.04 psu from 1960 to 1990. If we consider a salinity difference between inflowing and outflowing water of perhaps 2.0 psu, this change represents an increase of 2% over 30 years. Recently, much bigger changes have been reported in the eastern Mediterranean by *Roether et al.* [1996] and *Klein et al.* [1999], where the deep water formation site shifted to the Aegean Sea sometime between 1987 and 1995. There is now considerable debate over the extent to which the eastern basin salt content has actually increased and how much of the change represents salt redistribution from upper to lower levels. *Roether et al.* [1996] have estimated that the mean salinity of the eastern basin below 1500 m increased by about 0.05 psu over the 7 years 1987–1995, which would represent a 2–3% increase in deep water salinity relative to the inflow salinity over 7 years.

[36] We now turn to the changes in net $E-P-R$, which may have been responsible for some of these salinity changes. *Bethoux and Gentili* [1999] have shown that the net deficit in freshwater flux over the Mediterranean basin has increased considerably since the 1950s. They estimate an increase in excess evaporation of 7–10 cm yr^{-1}, due to changes in river inputs from the Nile, the Ebro, and other northern rivers and to decreases in precipitation over the area. This represents an increase in net evaporation on the order of 10%, much of which occurred several decades ago with the damming of major rivers such as the Nile.

[37] It seems clear that the increase in Mediterranean salinity detected up to 1990 was rather smaller in percentage terms than the $E-P-R$ change. This suggests that at least up to 1989, the period studied by [*Rohling and Bryden*, 1992], the basin may have responded to the net $E-P-R$ mainly by increasing the overturning. The small increases in salinity $\sim$0.04 psu are similar to the changes seen in our model experiments with 5–8% $E-P-R$ changes. The sudden changes in the eastern Mediterranean occurring around 1990 could signal a change in basin response toward producing more saline deep waters. If this were associated with increased mixing at Gibraltar, then there would be long-term consequences. If the basin deep waters become saline very quickly, it is more difficult to return to the previous state, and the deep basin is far more likely to become stagnant and anoxic. Recent evidence for regime shifts at Gibraltar are given by *Ross et al.* [2000], who suggest that the flow may have been closer to a maximal state from 1990 to 1992, from an analysis of sea level data either side of the straits, associated with the large increase in deep water formation within the basin. However, their analysis suggests the flow then returned to submaximal for part of each year after 1994.

[38] However, the basin might be responding to the flux changes directly by increasing salinity,and the only reason that the changes found by [*Rohling and Bryden*, 1992] are small in percentage terms could be due to the short time elapsed since the imposition of the $E-P-R$ change. In any case the radical changes in the model mixing regime at Gibraltar have not yet been observed in the real system. More observational monitoring of mixing at straits and of the water formation process itself and the air-sea fluxes which cause it is needed. Modeling work is also needed to understand better the range of model responses to changes in air-sea fluxes and to permit the application of observed air-sea fluxes to models without causing unrealistic model drifts.

6.2. Summary

[39] An OGCM of the Mediterranean is forced by surface fluxes of heat and freshwater. The baseline fluxes are obtained from a long restoring experiment that accurately simulates the circulation and hydrography of the basin. The diagnosed fluxes also compare favorably with a number of observational flux estimates at the

basin scale. The model circulation remains steady when forced by the diagnosed fluxes, with increased internal variability [*Myers and Haines*, 2000]. The diagnosed fluxes were then modified (in a sample basin-averaged sense) to simulate temporal changes to the excess evaporation over the basin.

[40] Under small changes in the fluxes (±8%) the basin behavior is linear, with increased (decreased) excess evaporation producing enhanced (weakened) overturning circulation with little change in large-scale hydrography. Intermediate water formation within the basin responds by producing more (less) water of a similar density, mainly by increasing the area of water formation. The salinity of the water formed does change at the formation sites, but subsequent mixing ensures that the Gibraltar salinities remain remarkably constant. Thus the increase (decrease) in the basin-averaged salinity is associated with an increase (decrease) in the amount of deep water and intermediate waters.

[41] For larger changes in the excess evaporation the basin's behavior is nonlinear. A 20% decrease in excess evaporation leads to a collapse of deep water formation, although this has little effect on the exchange properties at the Strait of Gibraltar since a vigorous intermediate cell remains. Larger decreases are examined by *Myers* [2002], where the paleoceanographic significance is discussed. Increases of $E - P$ of 10% or more also lead to very nonlinear behavior. A 25% increase in $E - P$ initially leads to increased deep water formation, but a hydraulic jump-like feature soon develops at Gibraltar, leading to strong mixing between inflowing and outflowing waters, reducing the ability of the basin to export salt. A runaway salinification occurs before the circulation finally collapses after $\sim$140 years, leaving only intermediate water formation and an unventilated saline abyssal layer. A more detailed examination of this nonlinear behavior and the results associated with the hydraulic jump and mixing at the Strait of Gibraltar are presently underway.

[42] Finally, we note that while the degree of mixing that develops in the modeled exchange flow at Gibraltar may be stronger than is realistic, any mixing or variations in mixing in this vitally important region will have consequences for the budgets of the basin as a whole. Changes in these mixing patterns, perhaps induced by changes to the basin forcing, could cause very nonlinear behavior in the long-term response of the entire Mediterranean system. This should be considered as a warning given ongoing climatic change and the large changes in river inputs that have already been introduced around the Mediterranean over recent decades.

[43] **Acknowledgments.** We would like to thank Eelco Rohling, Kevin Stratford, Stephan Matthiesen, and others in the CLIVAMP project for helpful discussions. We would also like to thank two anonymous reviewers for their insightful comments. This work was funded by the EU MAST MEDNET program under contract MAS3-CT98-0189 and also an NSERC operating grant awarded to PGM.

References

Bethoux, J.-P., Budgets of the Mediterranean Sea: Their dependence on the local climate and on characteristics of the Atlantic waters, *Oceanol. Acta*, *2*, 157–163, 1979.

Bethoux, J.-P., and B. Gentili, The Mediterranean-sea, coastal and deep sea signatures of climatic and enviromental-changes, *J. Mar. Syst.*, *7*, 383–394, 1996.

Bethoux, J.-P., and B. Gentili, Functioning of the Mediterranean Sea: Past and present changes related to freshwater input and climate changes, *J. Mar. Syst.*, *20*, 33–47, 1999.

Bethoux, J.-P., B. Gentili, J. Raumet, and D. Tailliez, Warming trend in the Western Mediterranean deep water, *Nature*, *347*, 660–662, 1990.

Brasseur, P., J. M. Beckers, J. M. Brankart, and R. Schoenauen, Seasonal temperature and salinity fields in the Mediterranean Sea: Climatological analyses of a historical data set, *Deep Sea Res., Part I*, *43*, 159–192, 1996.

Bryden, H. L., and T. H. Kinder, Steady two-layer exchange through the Strait of Gibraltar, *Deep Sea Res., Part A*, *38*, 5445–5463, 1991.

Candela, J., The Gibraltar Strait and its role in the dynamics of the Mediterranean Sea, *Dyn. Atmos. Oceans*, *15*, 267–299, 1991.

Dalziel, S. B., Two-layer hydraulics: A functional approach, *J. Fluid. Mech.*, *223*, 135–163, 1991.

Garrett, C., R. Outerbridge, and K. Thompson, Interannual variability in Mediterranean heat and bouyancy fluxoutes, *J. Clim.*, *6*, 900–910, 1993.

Gent, P. R., and J. C. McWilliams, Isopycnal mixing in ocean circulation models, *J. Phy. Oceanogr.*, *20*, 150–155, 1990.

Gilman, C., and C. Garrett, Heat flux parameterizations for the Mediterranean Sea: The role of atmospheric aerosols and constraints from the water budget, *J. Geophys. Res.*, *99*, 5119–5134, 1994.

Haines, K., and P. Wu, A modelling study of the thermohaline circulation of the Mediterranean Sea: Water formation and dispersal, *Oceanol. Acta*, *18*, 401–417, 1995.

Hopkins, T. S., The thermohaline forcing of the Gibraltar exchange, *J. Mar. Syst.*, *20*, 1–31, 1999.

Josey, S. A., E. C. Kent, and P. K. Taylor, Ocean-atmosphere heat, in *Momentum and Freshwater Flux Atlas*, *Southampton Oceanogl. Cent. Rep. 6*, 30 pp. + figures, Southampton, U. K., 1998.

Killworth, P. D., D. Stainforth, D. J. Webb, and S. M. Paterson, The development of a free-surface Bryan-Cox-Semtner ocean model, *J. Phys. Oceanogr.*, *21*, 1333–1348, 1991.

Klein, B., W. Roether, B. Manca, D. Bregant, V. Beitzel, V. Kovacevic, and A. Luchetta, The large deep water transient in the eastern Mediterranean, *Deep Sea Res., Part I*, *46*, 371–414, 1999.

Klein,, B.,W. Roether, G. Civitarese, M. Gacic, B. Manca, and M. d'Alcala, Is the Adriatic returning to dominate the production of Eastern Mediterranean Deep Water, *Prog. Oceanogr.*, *27*, 3377–3380, 2000.

Lascaratos, A., R. G. Williams , and E. Tragou, A mixed-layer study of the formation of Levantine Intermediate Water, *J. Geophys. Res.*, *98*, 14,793–14,749, 1993.

Lascaratos, A., W. Roether, K. Nittis, and B. Klein, Recent changes in deep water formation and spreading in the eastern Mediterranean, *Prog. Oceanogr.*, *44*, 5–35, 1999.

Leaman, K. D., and F. A. Schott, Hydrographic structure of the convective region in the Gulf of Lions: Winter 1987, *J. Phys. Oceanogr.*, *21*, 575–598, 1991.

Levitus, S., Climatological atlas ofthe world ocean, *NOAA Prof. Pap. 13*, 173 pp., U. S. Govt. Print. Off., Washington, D. C., 1982.

Macdonald, A., J. Candela, and H. L. Bryden, An estimate of the net heat transport through the Strait of Gibraltar, in *Seasonal and Interannual Variability of the Western Mediterranean Sea, Coastal Estuarine Stud.*, vol. 46, edited by P. E. la Violette, pp. 13–32, AGU, Washington, D. C., 1994.

Myers, P. G., Flux forced simulations of the paleo-circulation of the Mediterranean, *Paleoceanography,* 10.1029/2000PA000613, in press, 2002.

Myers, P. G., and K. Haines, Seasonal and interannual variability in a model of the Mediterranean under derived flux forcing, *J. Phys. Oceanogr.*, *30*, 1069–1082, 2000.

Myers, P. G., and E. J. Rohling, Modeling a 200-year interruption of the Holocene sapropel S_1, *Quat. Res.*, *53*, 98–104, 2000.

Myers, P. G., K. Haines, and S. Josey, On the importance of the choice of wind stress forcing to the modeling of the Mediterranean Sea circulation, *J. Geophys. Res.*, *103*, 15,729–15,749, 1998a.

Myers, P. G., K. Haines, and E. J. Rohling, Modeling the paleo-circulation of the Mediterranean: The Last Glacial Maximum and the Holocene with emphasis on the formation of sapropel S_1, *Paleoceanography*, *13*, 586–606, 1998b.

Rahmstrof, S., A fast and complete convection scheme for ocean models, *Ocean Modell.*, *101*, 9–11, 1993.

Roether, W., B. B. Manca, B. Klein, D. Bregant, D. Georgopoulos, V. Beitzel, V. Kovacevic, and A. Luchetta, Recent changes in eastern Mediterranean deep waters, *Science*, *271*, 333–335, 1996.

Rohling, E. J., and H. L. Bryden, Man-induced salinity and temperature increases in Western Mediterranean Deep Water, *J. Geophys. Res.*, *97*, 11,191–11,198, 1992.

Ross, T., C. Garrett, and P.-Y. Le Traon, Western Mediterranean sea-level rise: Changing exchange flow through the Strait of Gibraltar, *Geophys. Res. Lett.*, *27*, 2949–2952, 2000.

Send, U., J. Font, G. Krahmann, C. Millot, M. Rhein, and J. Tintore, Recent advances in observing the physical oceanography of the western Mediterranean, *Prog. Oceanogr.*, *44*, 37–64, 1999.

Sparnocchia, S., G. P. Gasparini, M. Astraldi, M. Borghini, and P. Pistek, Dynamics and mixing of the eastern Mediterranean outflow in the Tyrrhenian basin, *J. Mar. Syst.*, *20*, 301–317, 1999.

Stratford, K., A note on the implementation of flux-limited advection for tracers in an ocean circulation model, *J. Atmos. Oceanic Technol.*, *16*, 1284–1290, 2000.

Sur, H. I., E. Ozsoy, and U. Unluata, simultaneous deep and intermediate depth convection in the northern Levantine Sea, winter 1992, *Oceanol. Acta*, *16*, 33–43, 1993.

Thuburn, J., Multidimentional flux-limited advection schemes, *J. Comput. Phys.*, *123*, 74–83, 1996.

Webb, D. J., An ocean model code for array processor computers, *Tech. Rep. 324*, Inst. of Oceanogr. Sci., Wormley, U. K., 1993.

Wesson, J. C., and M. C. Gregg, Mixing at Camarinal Sill in the Strait of Gibraltar, *J. Geophys. Res.*, *99*, 9847–9878, 1994.

Wu, P., and K. Haines, Modelling the dispersal of Lavantine Intermediate Water and its role in Mediterranean Deep Water formation, *J. Geophys. Res.*, *101*, 6591–6607, 1996.

Wu, P., and K. Haines, The general circulation of the Mediterranean Sea from a 100-year simulation, *J. Geophys. Res.*, *103*, 1121–1135, 1998.

K. Haines, Environmental Systems Science Center (ESSC), Reading University, Reading RG6 6AL, UK. (kh@mail.nerc-essc.ac.uk)

P. G. Myers, Department of Earth and Atmospheric Sciences, University of Alberta, 1–26 Earth Sciences Building, Edmonton, Alberta, Canada T6G 2E3. (myers@wessex.eas.ualberta.ca)