Computational Efficiency and Accuracy of Methods for Asynchronously Coupling Atmosphere–Ocean Climate Models. Part I: Testing with a Mean Annual Model

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ABSTRACT

Using the mean annual, globally-averaged, coupled atmosphere–ocean energy balance model of Harvey and Schneider, the effect on the transient climate response to a step function solar constant increase using a variety of asynchronous coupling methods is investigated. In asychronous coupling, the atmosphere is integrated with fixed ocean temperature for a period τ_a, and the ocean is integrated for a longer period τ_0 with one of several possible atmospheric assumptions. The process is repeated until equilibrium occurs. The assumption of fixed atmospheric temperature during ocean integrations, used in some atmosphere–ocean GCMs, leads to a large energy conservation error and a very slow transient response compared to the synchronously coupled response. Assuming fixed heat fluxes or a fixed atmosphere–ocean temperature difference greatly improves the asychronous transient response, but large errors still remain. Assuming fixed turbulent fluxes subject to a time lag, or extrapolating the trend in the atmosphere–ocean temperature difference with a variable τ_0 gives very good globally-averaged results, but the former method is of questionable utility in a nonglobally-averaged model as it involves large compensating errors in individual heat flux terms.

1. Introduction

For over a decade, the response of the surface climate to external forcing (e.g., CO_2 increase) has been considered a problem whose credible solution would require well-tested, coupled, atmosphere–ocean general circulation models (A/O GCMs). In a pioneering study, Manabe and Bryan (1969) explored the equilibrium surface climate of an A/O GCM which was integrated for 10^3 simulated years. However, since the numerical time step in the OGCM is an order of magnitude longer than that of the AGCM, sychronous coupling between the atmospheric and oceanic submodels was ruled out on economic grounds. Instead, an *asynchronous* coupling scheme was devised whereby the AGCM is not explicitly integrated during several OGCM time steps. In essence, this scheme requires assumptions as to how the AGCM climate would have been evolving during the longer OGCM time integration steps had synchronous coupling been in force. Since the mass, momentum, and energy exchanges between AGCM and OGCM will evolve in time according to the respective evolving climates of each submodel, the noncalculation of AGCM variables for a block of time will cause errors in air–sea fluxes and thus lead to errors in their evolving climatic states—and perhaps their equilibrium states as well. However, investigations of the climate response used to date with asynchronously coupled A/O GCMs have been concerned only with the equilibrium climate response, so that large errors in the transient have been regarded as unimportant under the assumption that the models were transitive (see Lorenz, 1968; or Schneider and Gal Chen, 1973). In view of the importance of air–sea coupling for predicting the transient response on 10^1–10^3 year time scales of the climatic system to time-evolving external perturbations like CO_2 buildup, (e.g., see Hoffert *et al.,* 1980; Schneider and Thompson, 1981; Bryan *et al.,* 1982; Thompson and Schneider, 1982; Harvey and Schneider, 1985a), it is important to investigate how various asynchronous A/O GCM coupling schemes affect (i) the fidelity of A/O GCM transient climatic responses and (ii) the equilibrium climatic statistics given some scenario of external forcing. This issue was discussed to varying extents by Schlesinger (1979), Dickinson (1981) and Ramanathan (1981).

Here, we use the globally-averaged Box-Advection–Diffusion, atmosphere/ocean energy balance model of Harvey and Schneider (1985, subsequently referred to as HS1) to investigate the sensitivity of the model's transient air and ocean surface temperature responses to a step function external forcing as a function of various asychronous coupling schemes, including those used in, or similar to, several published A/O GCM studies (e.g., Manabe and Bryan, 1969; Manabe *et al.,* 1975, 1979; Washington *et al.,* 1980). Not only will we attempt to identify an asychronous scheme that closely reproduces the transient solution of a 10^3 year synchronous run with the HS1 global model, but we will

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also analyze the model's air–sea fluxes in order to understand better why different schemes behave as they do. The latter analysis will help to provide guidance as to which, if any, asynchronous coupling schemes between atmospheric and oceanic submodels may have reasonable utility in dynamical, three-dimensional A/O GCM calculations. In this study we will concentrate on annually-averaged, globally-averaged variables, as this is commensurate with the limited resolution of HS1. Indeed, such a high degree of aggregation has the advantage of making cause-and-effect linkages between various asynchronous schemes and model results clearer to understand. In a subsequent paper (Harvey, 1985), using a model with seasonal, hemispheric, and land–sea resolution (Harvey and Schneider, 1985b, later referred to as HS2), the additional complications of regionality and seasonality will be examined to extend our preliminary analyses of the viability of various asynchronous A/O GCM coupling schemes.

2. Model description and methodology

The HS1 model is shown in Fig. 1 for the convenience of the reader. Details of its construction and transient behavior can be found in HS1 (see also Hoffert *et al.,* 1980). However, in view of the importance of the atmospheric coupling between air over land and air over oceans to the evaluation of a globally-averaged equivalent mixed layer (EML) depth, the reader might want to consult HS2 for a discussion of the EML used here, even though HS2 involves more spatial resolution

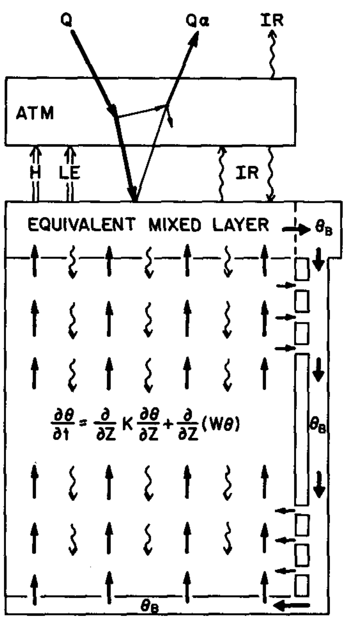

FIG. 1. The globally averaged mean annual model of Harvey and Schneider (1985a), used for the present study. *H*: sensible heat flux, *LE*: latent heat flux, IR: infrared flux, *Q*: incident solar radiation, α: planetary albedo, θ: ocean temperature, *K*: ocean thermal diffusion coefficient, and *w*: upwelling advection velocity.

than is treated here. Throughout this paper, we will investigate the response of HS1 to a step function 2% increase in solar forcing applied at time $t = 0$ as a function of various A/O coupling schemes, given in Table 1. We use as our reference the synchronously coupled transient response, which is energy conserving and uses equal time steps of one day for the atmosphere and ocean components.

Unless otherwise indicated, the atmosphere is integrated in the asynchronous mode with one-day time steps for a time τ_a with fixed ocean temperatures, followed by an ocean integration with 15-day time steps for a time τ_0 with one of the following atmospheric boundary conditions: fixed atmospheric temperature (fixed T_a); fixed solar, downward longwave, latent, and sensible heat fluxes (fixed flux); fixed atmosphere–ocean temperature difference (fixed ΔT); and fixed latent and sensible heat fluxes, but solar and infrared fluxes computed assuming fixed ΔT (fixed turbulent flux). For fixed ΔT and fixed turbulent flux, the atmosphere temperature is incremented by the amount of ocean warming during the previous ocean cycle before each atmosphere integration. The τ_a and τ_0 integrations are repeated until equilibrium is achieved. In most of the experiments that appear in Figs. 2–4, we use $\tau_a = 15$ days, as this corresponds to a reasonable atmospheric thermal relaxation time both in AGCMs and in our model (Washington and Chervin, 1980).

The assumption of fixed ΔT assumes that the atmosphere instantly follows the ocean warming during the ocean integrations. In reality the atmosphere would not respond instantly to the ocean warming but would experience a lag. We therefore vary the fixed ΔT and fixed flux assumption by introducing a lag τ in the atmospheric response at the beginning of each ocean integration. In practice we simply hold T_a fixed for a time τ, allowing ΔT to increase, then hold ΔT fixed at the higher value. The fixed T_a assumption was used by Manabe *et al.* (1979) and Washington *et al.* (1980) in seasonal A/O GCMs, whereas Manabe and Bryan (1969) and Manabe *et al.* (1975) used a fixed flux assumption in a mean annual A/O GCM.

3. Results

a. Transient temperature and heat flux response

Figure 2a compares the transient response to a step function 2% solar constant increase using fixed T_a with $\tau_a = 15$ days and $\tau_0 = 2\frac{1}{2}$ and 5 years. Figure 3 shows the evolution of the individual heat fluxes for this and other cases during the first two atmosphere integrations and the first ocean integration for $\tau_0 = 2\frac{1}{2}$ years. We chose to test $\tau_0 = 2\frac{1}{2}$ and 5 years because these values bracket the value $\tau_0 = 4$ years used by Washington *et al.* (1980). As seen in Fig. 3a, the upward sensible heat flux decreases with time for the synchronous case (solid line). This reduction is a result of a reduction in ΔT. The latent heat flux (Fig. 3b), on the other hand, in-

TABLE 1. Effect of various asynchronous coupling schemes on T_a and heat fluxes during ocean integrations for a 2% step-function increase of the solar constant.

Case	T_a (Implicit)*	Sensible heat flux	Latent heat flux	Downward IR flux	Incident solar flux
Fixed T_a	constant	increase	large increase	constant	constant
Fixed ΔT	increases	constant	increase	increase	small decrease
Fixed flux	—	constant	constant	constant	constant
Fixed turbulent flux	increases	constant	constant	increase	small decrease

* For cases in which T_a implicitly increases during ocean integration, we restart atmospheric integration with same ΔT as at the end of the previous atmospheric integration.

creases because of an increase in the surface-air vapor pressure difference which occurs in spite of the reduction in ΔT. Finally, as seen in Fig. 3c, there is a large increase in the downward longwave radiation in the synchronous case which, in absolute terms, is much larger than the changes in either the sensible or latent heat fluxes.

When the atmospheric temperature is held fixed for the 2½ year period after the initial 15 days of integration, the mixed layer temperature response is dramatically slowed. The cause of this is clear from Fig. 3. Because T_a is fixed while T_m is increasing, the sensible heat flux from the mixed layer increases in the fixed-T_a case, even though it would have decreased in synchronous mode. This is equivalent to a loss of energy from the ocean that, in essence, disappears from the system since T_a is not warming in response to the increase in upward sensible flux. Concurrently, downward IR radiation flux from the atmosphere to the mixed layer remains constant rather than increasing. This too represents a decrease in the perturbation flux to the mixed layer, which combined with the increased *upward* sensible heat flux, helps to explain the erroneously slow temperature response in the fixed-T_a mode. At the same time, the upward latent heat flux is slightly increased relative to the synchronous case for the early part of the asynchronous integration (see Fig. 3b), but lags behind for the later part. However, the large net loss of energy from the mixed layer (which is effectively lost from the entire climate system) seen in Fig. 3 dominates the signal and causes the very slow mixed layer response seen in Fig. 2a.

As a result of the large sensible and initial latent heat fluxes that arise with fixed T_a, there is very little further ocean response after the first few months of each ocean integration, so that there is very little ocean temperature change during most of the 2½ or 5 year ocean integrations, and most of this time is wasted. It may appear surprising that the ocean would equilibrate so quickly with the imposed fixed T_a boundary condition, given that the model ocean *e*-folding time is generally on the order of 10–20 years (Harvey, 1986). However, this *e*-folding time range applies to the coupled atmosphere–ocean system and is dependent on the system equilibrium sensitivity λ, which can be defined by $\Delta T(eq) = \lambda G$, where $\Delta T(eq)$ is the equilibrium temperature response and G is the initial forcing change. When the ocean is integrated asynchronously with fixed T_a, however, the effective sensitivity is greatly reduced, with a corresponding reduction in the ocean response time scale. As shown by Cess and Potter (1984), the surface temperature sensitivity when the atmosphere is allowed to respond is given by

$$\frac{1}{\lambda} = \frac{dF\uparrow}{dT_s} - \frac{dF\downarrow}{dT_s} + \frac{dH}{dT_s} + \frac{dLE}{dT_s} - \frac{dQ}{dT_s}, \quad (1)$$

where $F\uparrow$ and $F\downarrow$ are the upward and downward infrared fluxes at the surface, H and LE are sensible and latent heat fluxes, and Q is the incident solar flux. Our assumption of fixed T_a is comparable to that used by Newell and Dopplick (1979) to estimate the CO_2 induced equilibrium surface warming. As Cess and Potter (1984) point out, the sensitivity for this case is given by

$$\frac{1}{\lambda} = \frac{dF\uparrow}{dT_s} + \frac{\partial H}{\partial T_s} + \frac{\partial LE}{\partial T_s}, \quad (2)$$

which is considerably less than given by (1). Hence, with fixed T_a, there is a relatively rapid but small ocean response at the beginning of each ocean integration. This intermittent response is indicated for the first two τ_a and τ_0 cycles in Fig. 2a; thereafter, a smooth curve is drawn even though the response continues to be intermittent.

Perhaps some of these difficulties could be mitigated by using smaller values of τ_a and τ_0 than on Fig. 2a. Therefore, following Manabe and Bryan (1969), who used τ_a = ⅛ day and τ_0 = 12.5 days (but for fixed flux, which we try later),, we also choose a fixed T_a case with τ_a = ⅛ day and τ_0 = 12.5 days (in this case we use time steps of 1/16 day and 3.125 days for the atmosphere and ocean, respectively). Results for this case are also shown in Fig. 2a, and lead to some improvement compared to the τ_a = 15 days, τ_0 = 2.5 years case. However, as shown in Harvey (1985), when seasonality is in-

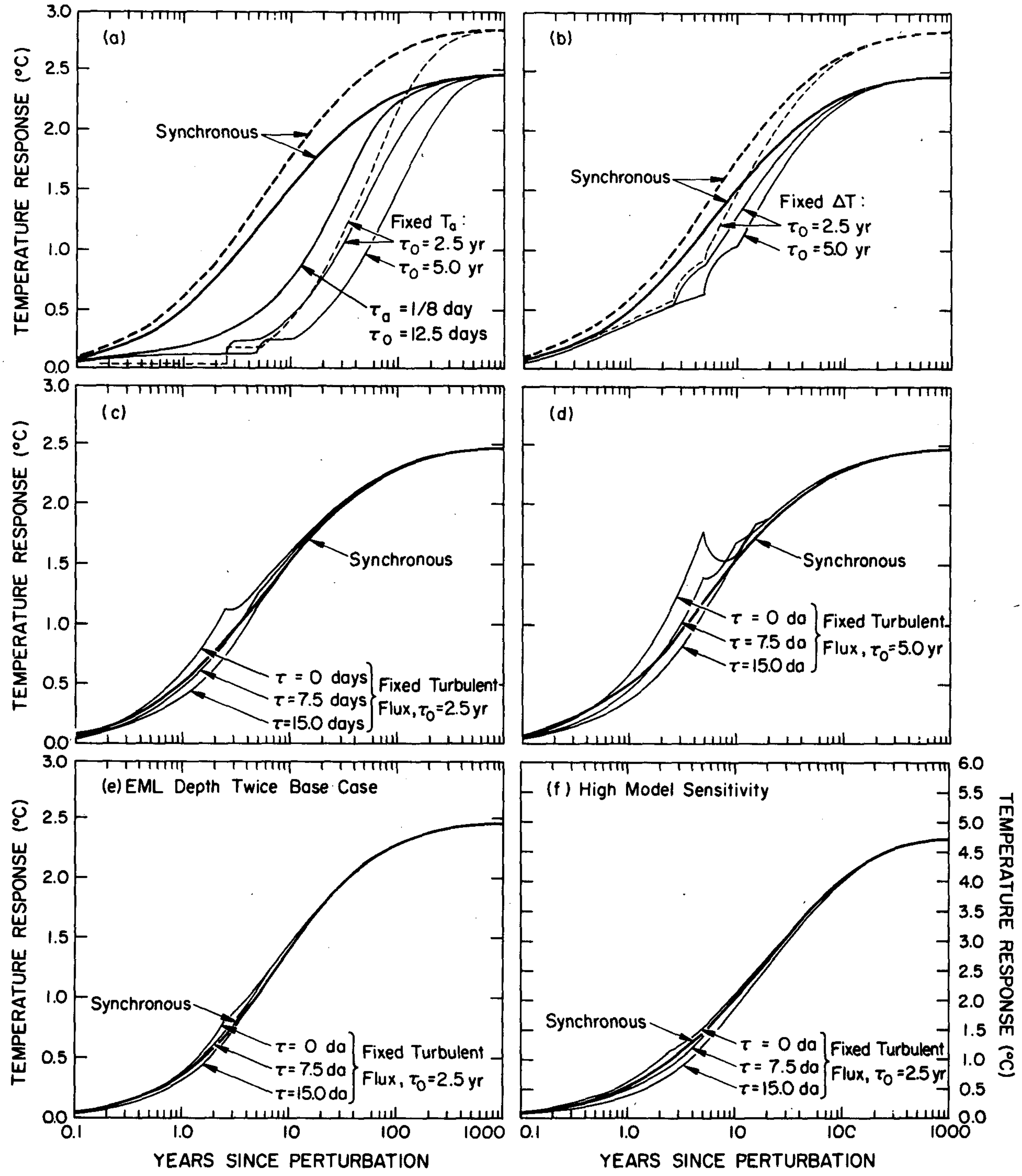

FIG. 2. Comparison of the atmospheric (dashed lines) and mixed layer (solid lines) transient temperature response for the synchronous case (heavy lines) and (a) fixed T_a, (b) fixed ΔT or fixed flux, (c) fixed turbulent flux with $\tau_0 = 2\frac{1}{2}$ years, base case model, (d) fixed turbulent flux with $\tau_0 = 5$ years, base case model, (e) as in (c) but for twice base case equivalent mixed layer depth, and (f) as in case (c) but for high model sensitivity with $B = 0.99$ W m^{-2} K^{-1}.

troduced (as in Manabe *et al.*, 1979), the transient response is considerably slower for this case. Moreover, although reducing τ_a and τ_0 improves the results in the global energy balance model we use (HS1), in a 3-D dynamical model that internally generates weather variability, the reduction of τ_a to less than a few days

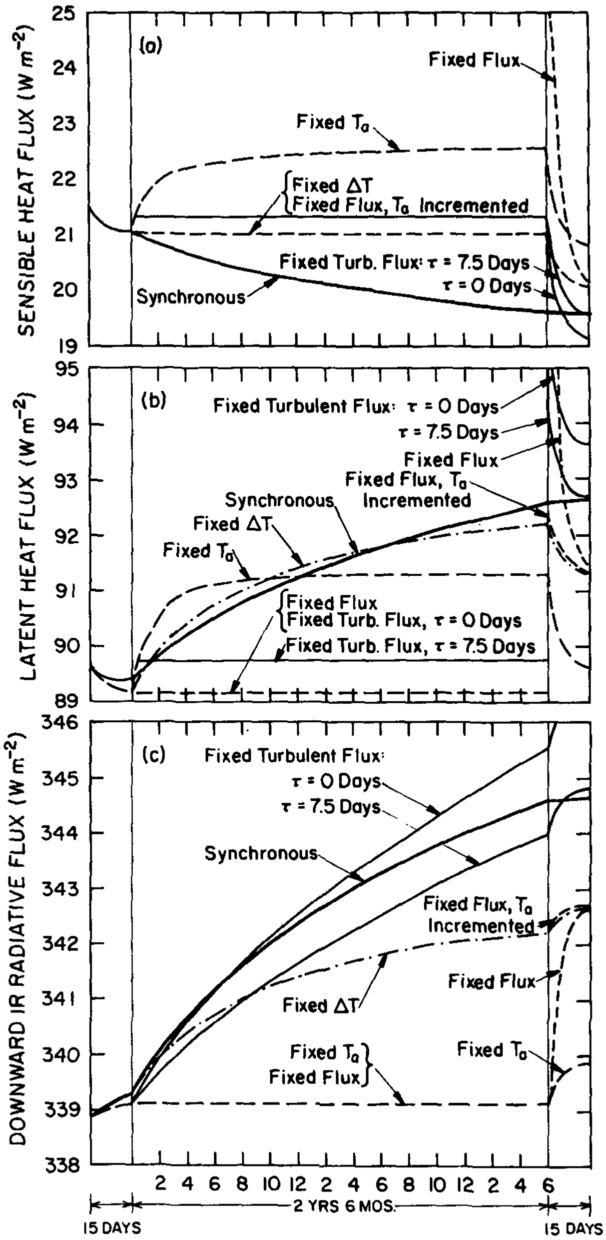

FIG. 3. Comparison of heat fluxes for the synchronous case and various asynchronous cases during the first two atmospheric integrations and an ocean integration of 2½ years. Note the variable horizontal scale. (a) Sensible heat flux, (b) latent heat flux, (c) downward IR radiative flux.

can cause substantial sampling problems associated with the extrapolation of fluctuating local conditions over an ocean interaction period τ_0. Thus, what may be gained by reducing τ_a could be lost locally by sampling errors (see also Harvey, 1985).

An obvious way to resolve the perturbation energy loss of the fixed T_a scheme is to assume that the atmosphere warms up instantaneously at exactly the same rate the mixed layer does during the asynchronous ocean integrations, as in the fixed ΔT scheme described earlier. Note that, as Fig. 2b shows, the fixed ΔT transient response is much closer to the synchronous case, although it is still slower and the rate of warming within each asynchronous cycle is still nonuniform. As shown in Fig. 3, the sensible heat flux remains constant during each ocean integration, rather than increasing dramatically as for fixed T_a, and the initial increase in the latent heat flux is much smaller. Thus, less heat is lost from the mixed layer as it warms. At the same time, the downward IR radiative flux increases, but by less than for the synchronous case. The net result is a transient mixed layer response that is much faster than for fixed T_a, but is still slower than for the synchronous case.

We next consider the fixed flux case, which is similar to that used by Manabe and Bryan (1969) and Manabe *et al.* (1975) in a mean annual model. In this case the upward sensible and latent heat fluxes do not increase as the mixed layer warms, but neither does the downward IR increase (Fig. 3). The net effect for τ_a = 15 days and τ_0 = 2.5 or 5 is a mixed layer transient response which is almost indistinguishable from that for fixed ΔT shown in Fig. 2b, so we do not show these results separately. We also try the case where τ_a = ⅛ day and τ_0 = 12.5 days, used by Manabe and Bryan (1969) and Manabe *et al.* (1975). In this case the transient response is almost indistinguishable from the fixed T_a, τ_a = ⅛ days, τ_0 = 12.5 days case shown in Fig. 2a. It would appear that with such short atmospheric integration times the change in T_a or in fluxes is not large during each atmosphere–ocean iteration, so that it does not matter what assumption one makes. As with fixed T_a, the fixed flux case does worse in a seasonal model (Harvey, 1985).

Note that in the synchronous case the atmosphere warms up more than the mixed layer (see Fig. 2a), thereby explaining why the downward IR flux into the mixed layer is underestimated and the upward turbulent fluxes overestimated in the fixed ΔT case with τ_a = 15 days. To partly remedy this situation, we derived a "fixed turbulent flux" case in which sensible and latent heat fluxes were held fixed during the ocean integration as for fixed flux, but downward IR radiative flux was increased as for fixed ΔT. In the synchronous case the total turbulent flux increases as the ocean warms (but by less than for the fixed ΔT case). The total turbulent flux in the fixed flux case is thus underestimated relative to the synchronous case (for a warming response), but the turbulent flux error almost exactly cancels the downward IR error. Indeed, with fixed turbulent flux, we overestimate the ocean heating, resulting in a mixed layer transient response that is slightly too fast. However, if we assume a lag of length τ in the atmospheric response to the oceanic surface temperature increase, during which time ΔT increases, we can improve the transient response. As seen in Fig. 2c, the fixed turbulent flux cases produce surprisingly good transient temperature simulations, particularly when $\tau = 7.5$ days. Inasmuch as τ represents the effect of the atmospheric thermal inertia, and a week or two is a reasonable radiative relaxation time for the atmosphere, perhaps it should not be surprising that choosing $\tau = 7.5$ days gives good results. However, we wish to emphasize that the good performance of the fixed turbulent case with nonzero τ arises partly from compensating errors in the individual heat fluxes. The degree of compensation is dependent in particular on the choice of τ_0 and τ. This fact is illustrated in Fig. 2d, which shows the transient mixed layer response for fixed turbulent flux when $\tau_0 = 5$ years. For $\tau = 7.5$ days, the mixed layer response is slightly less than that of the synchronous case during the first 2½ years, but is much too large after 5 years. When $\tau_0 = 2½$ it just happens that we recalibrate all the fluxes with an atmosphere integration when the mixed layer temperature is almost identical to what it was for the synchronous case. If τ_0 is longer, we need a longer τ in order to get near compensation of the individual flux errors.

The final two cases (Fig. 2e and 2f) are cross-sensitivity experiments (see HS1), whereby we investigate the transient response to our generic step-function solar constant perturbation for synchronous and $\tau_0 = 2.5$ years fixed turbulent flux asynchronous cases in which the equivalent mixed layer depth is doubled (Fig. 2e) and the model's equilibrium sensitivity is doubled (Fig. 2f) by halving our radiation damping coefficient (see HS1). In view of the importance of these parameters to transient climatic simulations (see HS2), we show these cases here. Both Fig. 2e and 2f show similar qualitative behavior to Fig. 2c, but the relative errors between synchronous and asynchronous transient temperatures are worse in the case of Fig. 2c. This suggests that quantitative determination of the reliability of various asynchronous coupling schemes will generally depend on those model parameters which influence model response characteristics.

As with the base case model parameters, the fixed turbulent flux cases shown in Fig. 2e and 2f involve large individual flux errors, even though the globally-averaged transient temperature response is very close to that of the synchronous case. When the fixed turbulent flux assumption is implemented in a seasonal model, however, the result is disastrous, with the transient response frequently becoming completely unstable (Harvey, 1985). Our only reason for showing the fixed turbulent flux results here is to make the generic point that an asynchronous coupling scheme that works well at one time or space scale may not work well at other scales if it works well at one scale only because of the fortuitous compensation of large individual heat flux errors. A criterion for robustness in an asynchronous coupling scheme is not only that it closely reproduce the synchronous response for a variety of model parameters (such as model sensitivity and mixed layer depth), but that it also correctly reproduces the time evolution of individual heat fluxes.

To develop a more robust asynchronous coupling scheme, we abandon our use of fixed τ_0 and instead adopt a strategy using variable τ_0. We start with small τ_0 at the beginning of the transient response and gradually increase it to some final value, which is then maintained to the end of the simulation. We represent the sequence of τ_0 by a series $\{\tau_1, \tau_2, \tau_3, \ldots, \tau_N\}$, where successive terms give the value of τ_0 for successive ocean integrations, and the last term is the final τ_0 used for the remainder of the simulation. Among the atmospheric boundary condition assumptions examined so far, we consider only fixed ΔT for further analysis, as it is seemingly the most physically realistic. As already noted, however, the atmosphere–ocean temperature difference decreases as the system warms, rather than remaining constant. We therefore also test an extrapolated ΔT scheme, whereby ΔT is estimated using a second-order Taylor series extrapolation based on the latest three ΔT values, as determined from the atmospheric integrations. Reliable extrapolation is not possible during the first ocean integration, and during the second ocean integration we extrapolate ΔT linearly, so that full second-order extrapolation does not begin until the third ocean integration. Furthermore, because the atmosphere initially responds much faster than the mixed layer, resulting in a rapid initial reduction in ΔT, the initial τ_0 must be long enough that we can reliably estimate $d\Delta T/dt$ and $d^2\Delta T/dt^2$ for use during subsequent longer ocean integrations.

Figure 4 compares the synchronous, fixed ΔT, and extrapolated ΔT transient mixed layer responses for $\tau_0 = \{½, ½, 1½, 2½\}$ years and $\tau_0 = \{½, ½, 1½, 2½, 5\}$ years, both of which satisfy the aforementioned requirement for extrapolated ΔT of gradually increasing τ_0. The temperature errors even for fixed ΔT are comparable to those for fixed turbulent flux with $\tau_0 = 2½$ years, and are even smaller for extrapolated ΔT. More importantly, the errors in the individual heat fluxes for extrapolated ΔT are all less than 0.1 W m^{-2} after the first ½ year asynchronous ocean integration, suggesting

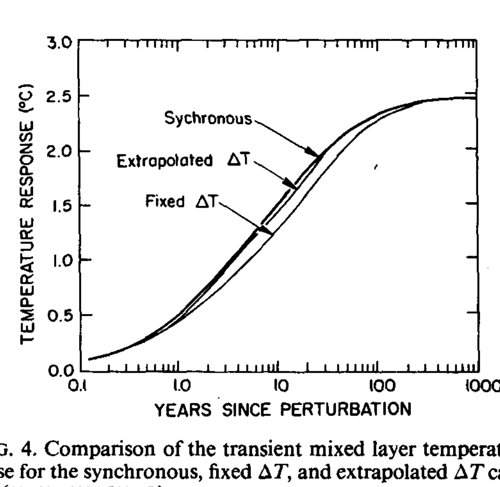

FIG. 4. Comparison of the transient mixed layer temperature response for the synchronous, fixed ΔT, and extrapolated ΔT cases for $\tau_0 = \{½, ½, 1½, 2½, 5\}$ years.

that this method may be robust, even in a A/O GCM—a suggestion that is supported further by results obtained with a model having a seasonal cycle and land-sea resolution (Harvey, 1985).

b. *Energy conservation errors*

All of the asynchronous coupling schemes previously discussed fail to some degree to conserve energy. The energy conservation error is, of course, largest for fixed T_a, where the large sensible and latent heat fluxes that develop are effectively lost to space since the atmosphere is not allowed to warm. To measure the degree of energy nonconservation for various asynchronous coupling cases, we compare the total change in the heat content of the system at the end of the integration with the integrated net heat flux at the top of the atmosphere. For the synchronous case the relative energy conservation error is on the order of 10^{-7}. For the asynchronous cases we compute the net heat flux during ocean integrations using the same T_a as is assumed in computing the net ocean heating. Inasmuch as a unique T_a is not defined for fixed flux, we show in Table 2 the relative energy conservation errors only for the fixed T_a, fixed turbulent flux, fixed ΔT, and extrapolated ΔT cases. The energy errors are negative, implying that the change in atmospheric and oceanic heat content is smaller than the total energy input. As seen from Table 2 and Fig. 2, the energy errors are proportional to the errors in the transient temperature response. It would appear, therefore, that the computed energy conservation error can serve as a useful diagnostic tool of the overall performance of a given asynchronous scheme. However, as shown in Harvey (1985), as soon as land-sea resolution is introduced the computed energy conservation error can no longer be a reliable indicator of the performance of a given asynchronous coupling scheme. Furthermore, calculation of the energy conservation error for A/O GCMs would entail an expensive recomputation of additional atmospheric radiative heat fluxes during the ocean integrations, thereby reducing the efficiency of asynchronous coupling. Thus, we show the energy conservation error here largely because of its conceptual value in illustrating the differences between various asynchronous coupling schemes.

TABLE 2. Percent energy conservation error associated with various asynchronous coupling methods. Negative errors imply that the change in atmospheric and oceanic heat content is smaller than the net energy input for a 2% step-function increase of the solar constant.

	Case	Error
Fixed T_a	τ_a = 15 days, τ_0 = 2½ yr	−61.7%
	τ_0 = 5 yr	−76.8%
Fixed ΔT	τ_0 = 2½ yr	
	τ_0 = 5 yr	−21.6%
	τ_0 = {½, ½, 1½, 2½} yr	−9.8%
	τ_0 = {½, ½, 1½, 2½, 5} yr	−15.9%
Fixed turbulent flux	τ_0 = 2½ yr no lag	−3.3%
	2½ yr, 7.5 day lag	−0.9%
	2½ yr, 15 day lag	−1.9%
	τ_0 = 5 yr, no lag	−9.3%
	5 yr, 7.5 day lag	−6.4%
	5 „ 15 day lag	−3.0%
Extrapolated ΔT	τ_0 = {½, ½, 1½, 2½} yr	−4.1%
	τ_0 = {½, ½, 1½, 2½, 5} yr	−6.5%

4. Concluding comments

We have shown that it is possible to experiment with various asynchronous coupling schemes between atmospheric and oceanic submodels in order to obtain a fairly good asynchronous simulation of the synchronous transient surface temperature response to a step function perturbation in external solar forcing. Of course, this is done in the context of a globally-averaged, annually-averaged, energy balance atmosphere/ocean model. Although we have, in particular, identified two different methods which both give very good results, in one of these methods (fixed turbulent flux with constant τ_0) this good performance occurs as a result of large compensating errors in individual heat flux terms.

Our results serve to illustrate some fundamental ideas involved in asynchronous coupling, but many questions remain concerning the practical implementation of these ideas in complex, three-dimensional A/O GCMs. These questions include the effect of a seasonal cycle, land-sea resolution, stochastic variability (especially locally in an AGCM), and spatially and temporally inhomogeneous feedback processes such as those involving snow and sea ice. With our model we were able—with no economic hardship—to compute the synchronous "truth," against which we compared various asynchronous coupling methods. Obviously, deriving the analogous synchronous "truth" for 10^3 years for an A/O GCM is presently beyond most practical computing budgets. As an alternative, we propose testing asynchronous coupling methods using a hierarchy of simple models of increasing spatial resolution and physical complexity, but still simple enough to permit computation of the synchronously coupled transient response. This process, which we have begun here, is continued by Harvey (1985), using a model having land-sea resolution and a seasonal cycle.

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