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CHAPTER 9

TRANSIENT EFFECTS DUE TO OCEANIC THERMAL INERTIA IN AN ATMOSPHERIC MODEL COUPLED TO TWO OCEANS

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ABSTRACT

A model of the mixed layers of the North Pacific and North Atlantic Oceans has been coupled to a hemispherical, quasigeostrophic, beta-plane model of the atmosphere. The two oceans are represented by the sectors 90°W–90°E and 0°–45°W, respectively, with the North American and Eurasian land masses by 45°–90°W and 0°–90°E. The mixed-layer model is a simplification of that of Kraus and Turner (1967). The meridional heat flux in each ocean is prescribed during each run of the model and several experiments were carried out between which these transports were varied. This paper considers the role of oceanic thermal inertia in these experiments. Each experiment began with a mixed layer which was everywhere in dynamic equilibrium with the unperturbed heat transport. The results show how the mixed layer affects the dynamics of the model. There are teleconnections between the mixed layers of the two oceans with time scales of months. The model shows an east–west oscillation reminiscent of the seesaw in northern hemisphere temperatures described by Van Loon and Rogers (1978), which involves the northernmost mixed layer of the oceans.

INTRODUCTION

Manabe (1983) highlighted three ways in which the oceans affect the climate, via: (1) the hydrological cycle; (2) their thermal inertia; and (3) the transport of heat. The first of these is not considered in the present model. Murdoch and Taylor (in prep.) have investigated the influence on the climate of the meridional oceanic heat flux using a hemispherical, quasigeostrophic, beta-plane model of the atmosphere which is coupled to two land masses and two oceans, each ocean having an interactive mixed layer. The transport of heat within each ocean was prescribed throughout any experiment so that its impact on the atmospheric circulation could be determined. The experiments showed that: changes in the surface temperature spanning the whole northern hemisphere accompanied perturbations of the oceanic heat fluxes; zonal wind strengths were inversely related to the strength of the heat transport, the wind changes being generally such as to oppose any change in the heat flux but with effects in both oceans; and the regions of deep convection in the northern Atlantic were important during the readjustment between climatic phases. These results were obtained by analysing the average conditions during the last 150 days of each 450-day run and assumed that the averages provided an indication of the equilibrium towards which the model was moving. The present paper examines the role of the thermal inertia of the oceans in the dynamics of the climatic system by studying the behaviour of the mixed layer during the evolution with time of this model.

The model uses an unchanging annual-average heating function throughout the calculations so that the heat content of the mixed layer is approximately constant. The real mixed layer will adjust to seasonal changes and so the response times of the thermal and potential energy contents of the layer to external forcing must be less than six months and may be similar to the decay time of sea-surface temperature anomalies (about two months, Frankignoul, 1979; Wells, 1982). The atmospheric index cycle has a period of about 20 days so that some response of the mixed layer to these atmospheric fluctuations can be anticipated. Therefore, the mixed layer may be expected to affect the dynamics of the atmosphere over monthly periods, and this is reflected in the large-scale transients to be described. The transients that occur exhibit zonal patterns which are similar to features observed in climatic data.

MODEL DESCRIPTION

The complete model is described in detail by Murdoch and Taylor (in prep.). The atmospheric part, which is that of Gordon and Davies (1977), is a two-level, quasi-geostrophic, beta-plane system, incorporating a model-dependent surface energy balance equation and, at the 500 mb level, a time-dependent heating function with associated feedback relationships due to radiative exchanges with the surface. The surface conditions are interactive; the albedo, which represents snow and ice cover, being a function of temperature. When the temperature at any point is below 272K the local albedo is changed to that of ice and if the area is sea it is treated as land. Radiation transmission parameters, sensible heat loss, latent heat loss and condensation heating were from Smagorinsky (1963); sensible and latent heat fluxes were allowed to vary linearly with surface temperatures about these mean values.

Surface temperatures over land, T_*, were calculated by assuming thermal equilibrium. Over the sea this is not permissible because heat can be stored in, or released from, the water column. The temperature, T_*, and depth, h, of each ocean's surface mixed layer were calculated by the method of Horgan, Davies and Gadian (1983) which is a simplification of that of Kraus and Turner (1967). Water temperature beneath the mixed layer is assumed to be a constant function of latitude (T_h). T_* and h were determined by ensuring that the thermal and potential energies of the layer were conserved. The atmospheric thermal and potential energy forcing functions are written as $S + B - M$ and $G - D + S/\gamma$, respectively, in which: S represents the contribution from solar heating; B the sum of all black-body, latent heat and sensible heat losses from the sea surface; G the rate of generation of mixing energy by the surface winds (deduced from the atmospheric model fields); S/γ the generation of buoyancy by the penetrative solar radiation (γ is the coefficient of decay of S with depth) and D the rate of dissipation of energy within the layer. Following Stevenson (1979), D was taken proportional to the depth of the layer, thereby ensuring no irreversible accumulation of potential energy. The term M is the oceanic heat-flux divergence which represents the meridional transport of heat within each ocean. Zonally averaged values of M (from Smagorinsky, 1963) were used and variations in the oceanic heat flux were implemented by changing M.

EXPERIMENTS

The model (Fig. 1a) partitions the Northern Hemisphere into four sectors: Pacific (90°E–90°W), North America (90°–45°W), Atlantic (45°W–0°) and Eurasia

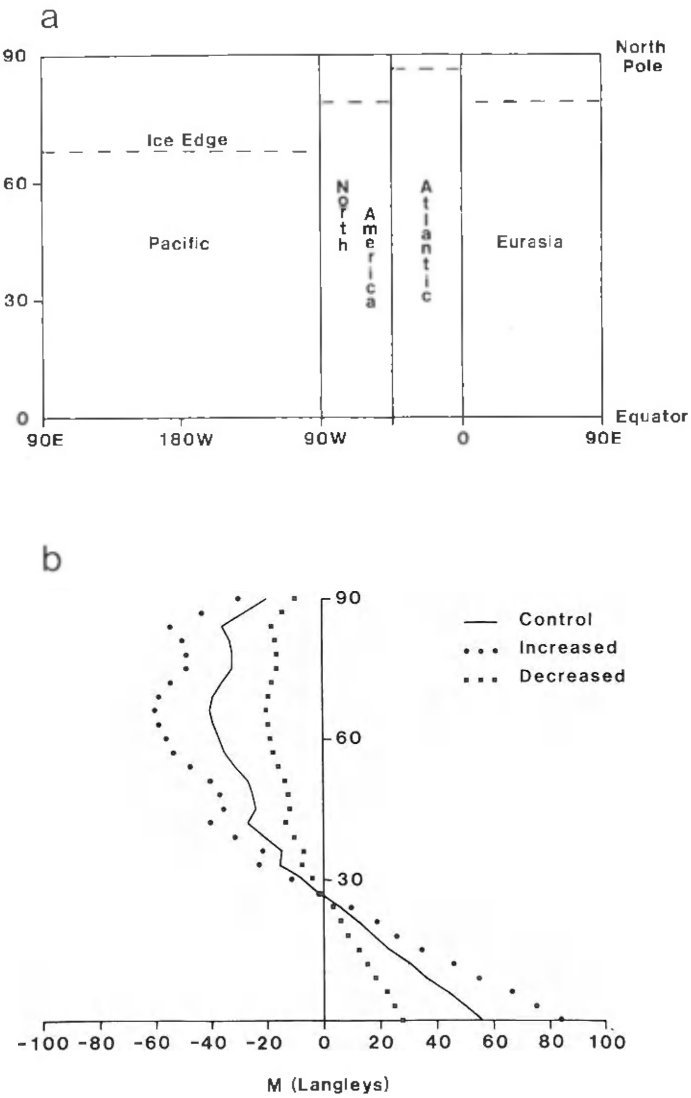


Fig. 1. a. The geometry used in the model. The extent of sea- and land-ice for the control run are shown. b. The latitudinal distribution of heat-flux divergence used in each ocean during the control run and during the runs with increased M and decreased M. Each unit of M represents 1 Langley in the Pacific or 4 Langleys in the Atlantic.

($0°$–$90°$E). The grid-spacing is $7.5°$ of longitude and $2.9°$ of latitude. There is no topography.

The experiments were designed to investigate the sensitivity of the system to changes in the oceanic heat-flux divergence (M) occurring in either ocean. In this model, M is prescribed in any run. There is considerable uncertainty about the relative strengths of the zonally averaged oceanic heat flux divergence in the two oceans. Meridional heat transports in the North Pacific and North Atlantic Oceans are similar, despite the greater area of the Pacific (Bryan, 1982): partly because there is a larger return flow of heat in the Pacific and partly because of the wide connection between the Atlantic and Arctic Oceans. Therefore, in the model, M was set up to be four times as great in the Atlantic sector as in the Pacific sector so that the two model oceans transported the same amount of heat northwards. The divergence was distributed uniformly with longitude. This arrangement was used for the control run.

The initial values of the stream functions were obtained by running the model without a mixed layer until the atmosphere came into equilibrium with the specified climatological heating (the zonally averaged heating functions of Smagorinsky, 1963). Subsequent initialisation of the mixed layer was carried out using the surface temperature and surface wind fields, averaged over several energy index cycles. By requiring that thermal and potential-energy forcing functions be simultaneously zero, that is:

$$S + B - M = 0 \tag{1}$$

and:

$$G + D - S/\gamma = 0 \tag{2}$$

these average fields could be used to provide initial oceanic-surface temperatures and mixed-layer depths. The first control run was obtained from the succeeding 450 days after the mixed layer became active. The calculation without a mixed layer was continued as a second control run.

Four experiments with differing M-values were carried out (Fig. 1b): (A1) M was increased by 50% in the Atlantic sector; (A2) M was decreased by 50% in the Atlantic sector; (P1) M was increased by 50% in the Pacific sector; and (P2) M was decreased by 50% in the Pacific sector.

Each of these experiments started with the same initial mixed layer and atmosphere as did the first control run and ran for at least 450 days. Comparing the pair of opposite runs A1 and A2 (or P1 and P2) to see which responses reverse sign provides an indication of signal to noise ratio.

RESULTS

Figure 2 shows the temporal variation of the atmospheric eddy kinetic energy for the control run with and without a mixed layer. When the oceans have a mixed layer the kinetic energy varies more smoothly with time, and this is consistent with analyses of models under stochastic forcing (e.g., Frankignoul, 1979) in which the thermal inertia of the ocean transforms a white noise spectrum of sea-surface temperature variations into a spectrum with greater energy in the low-frequency regime. However, Fig. 2 shows this reddening to be more complex, for peaks at 30 days and 75–100 days are smeared into a broad band centred on a period of 50 days.

The presence of oceanic mixed layers in the model leads to a generally warmer climate (Murdoch and Taylor, in prep.). For instance, although ice cover on land does not change appreciably, the extent of sea-ice is reduced by about six degrees of latitude. This warming can be understood if the development in time of the model is examined in the

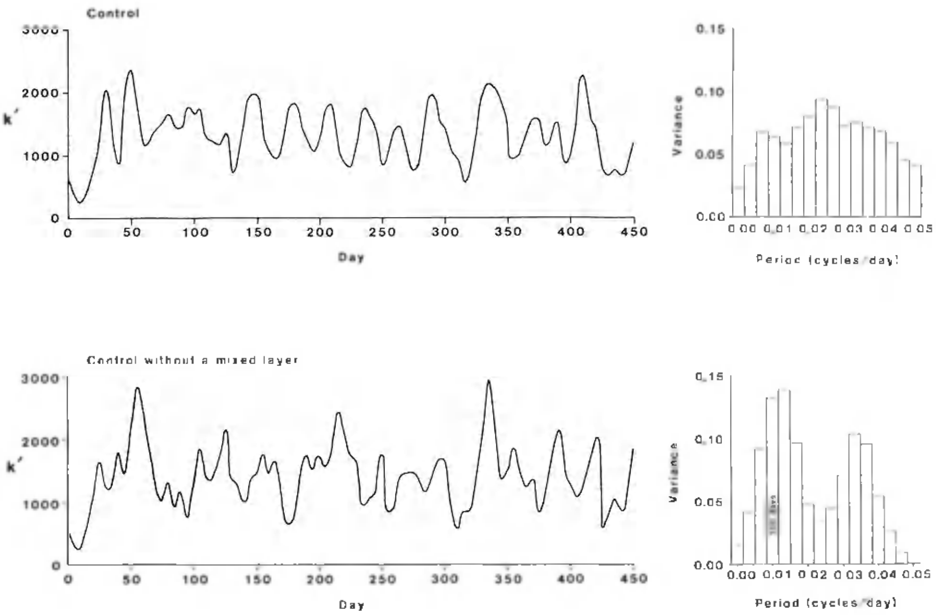


Fig. 2. Time series of mean eddy kinetic energy for the control run with and without a mixed layer. The low-frequency spectrum is shown for each series.

light of the atmospheric dynamics and thermodynamics. The initial stages of Fig. 2 illustrate the way this difference in climate arises. Both time series show a trough at about day 10, after which each graph climbs to a peak. The series from the experiment with a mixed layer has a much higher first peak. This occurs because the presence of a heat-absorbing mixed layer requires that more heat be transported northwards by the atmospheric eddies to produce a given temperature change. So, more vigorous eddies can build up before the temperature gradient at the 500 mb level is forced back towards baroclinic stability. During this period, heat that would otherwise be lost by radiation to space is stored in the mixed layer. Such a storage of heat occurs during each of the atmosphere's energetic phases and this heat can be released at less energetic times to impede the formation of ice. Thus, the mixed layer affects the climate of the model by influencing the atmospheric thermodynamics via the 500-mb temperatures and hence the dynamics of the atmosphere through the meridional temperature gradient. The mixed layer is therefore seen to have a significant effect on the system even in the absence of seasonal forcing.

The fluctuations shown in Fig. 2 are dominated by the atmospheric energy cycle. However, the experiments A1, A2, P1 and P2 having modified oceanic heat fluxes each began with mixed layers which were not in "equilibrium" since, due to the altered heat transports, their thermal and potential-energy forcing functions were not zero. As a result the system gradually adjusts during each run towards conditions which satisfy eqns. (1) and (2) and fluctuations with longer time scales are also observed. These fluctuations can be seen in Fig. 3 which displays for two typical cases, A1 and A2, successive 50-day averages of the heat stored in the mixed layer (after subtracting that of the control run) at a southern latitude and at a latitude near the northernmost exposed region of each ocean. For the whole of the Pacific and the southern part of the Atlantic the changes are slightly oscillatory and it is possible that an equilibrium is being approached. In the northern Atlantic, however, a clear trend is apparent. Mixed-layer depths in the two oceans (Fig. 3) also show less trend at low latitudes but, at high latitudes, the depths show a trend which has the same sign in both oceans. All of these trends change sign when the perturbation in oceanic heat flux is reversed. The mixed layers at high latitudes will take longer to achieve a dynamic equilibrium than the layers at low latitudes because they are much deeper and so require the transfer of a considerably larger quantity of heat to cause a given temperature change.

At low latitudes, the trend in the heat content of the mixed layer is the same in both oceans and corresponds to the underlying trend in the mixed-layer depth (Figs. 3 and 4). The winds in the model respond rapidly (i.e. within 50 days) to the changing oceanic heat flux and the trend in the layer depth represents a slower adjustment to these winds. At high latitudes this is also true of the Pacific in runs A1, A2 (Fig. 3) and the Atlantic in P1, P2 (e.g., Fig. 4). Thus, there is a teleconnection between the two oceans affecting most latitudes. The northern mixed layer in the ocean whose heat flux is perturbed shows a trend in its heat content which is the direct result of the changed heat transport and is in the opposite direction to that of the other ocean.

Superimposed on this overall trend, the mixed-layer depths in the northern region of the ocean show an oscillation of period about 400 days which is out of phase between the two oceans, thereby having the appearance of an east–west seesaw. The seesaw has its greatest amplitude in the high-latitude area of each ocean. As a result, the response in the Atlantic decreases towards middle latitudes and at 57°N in this ocean the oscillation is less clear. This transient response of the model is examined further in Fig. 4 which contains the same graphs for experiment P1 together with the corresponding time series of surface temperatures and zonal-wind components. They are typical of the four runs. At low latitudes where the mixed layers are shallow the fluctuations in surface temperature are relatively large. The temperatures and zonal winds appear inversely related to the layer depths indicating the importance of variations in wind mixing. In this "trade-wind" region, the winds are predominantly zonal easterlies, thus a positive zonal wind difference means a reduced wind strength. There is a similar set of relationships in high latitudes but, because of the trends in the layer depth and storage, the temperatures, and especially the winds, are less clearly related to the layer depths. The interpretation is also complicated because 57°N is in a band of westerlies, while 75°N is in a band of easterlies.

The mixed layers of the two oceans are coupled by means of the zonal winds which transmit changes over one ocean downstream to the other. Equations describing this ocean–ocean interaction can be obtained from those describing the thermal and potential-energy balances of the mixed layer, viz:

$$\frac{\partial}{\partial t}\,[(T_* - T_h)h] = S + B - M \tag{2}$$

where T_h is the temperature below the surface layer, and:

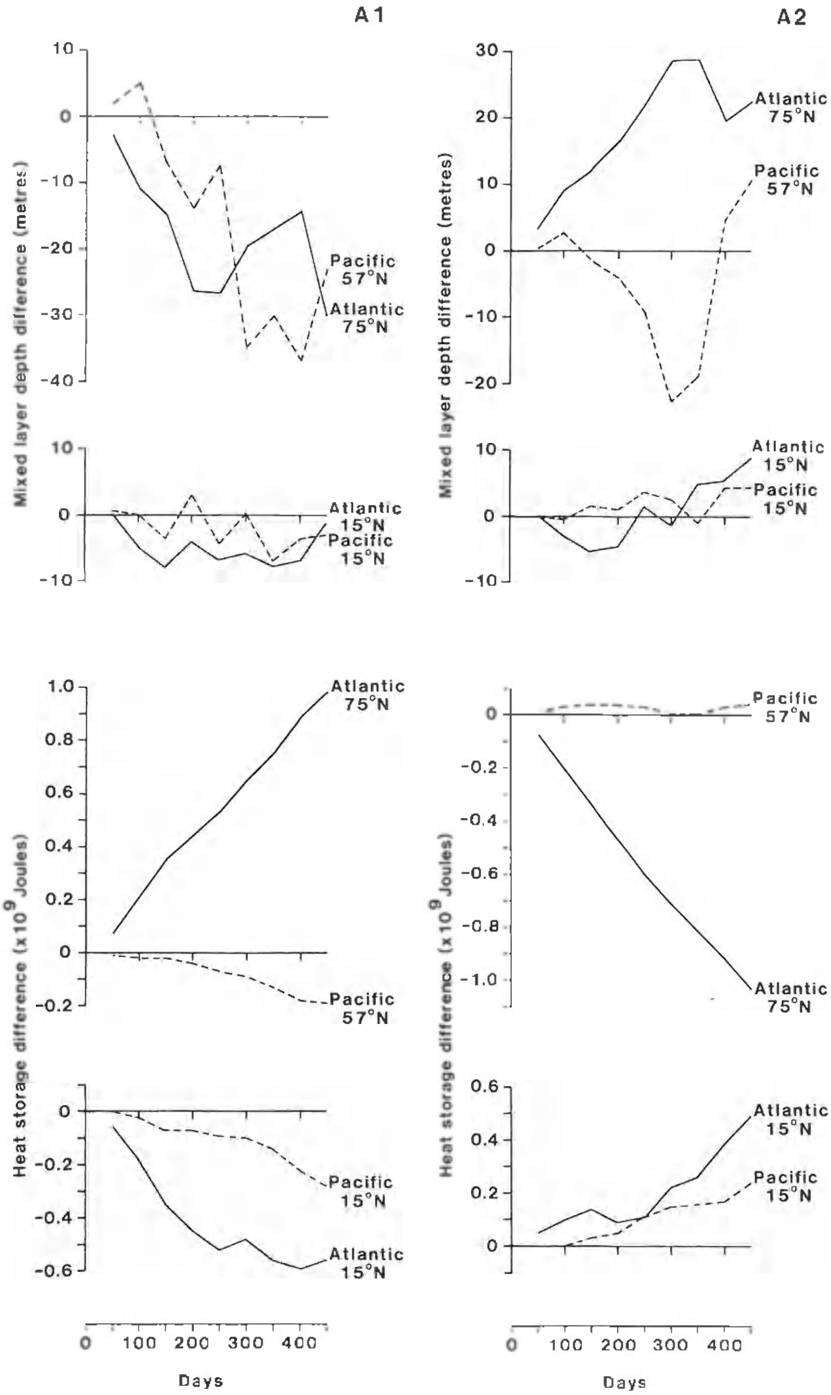
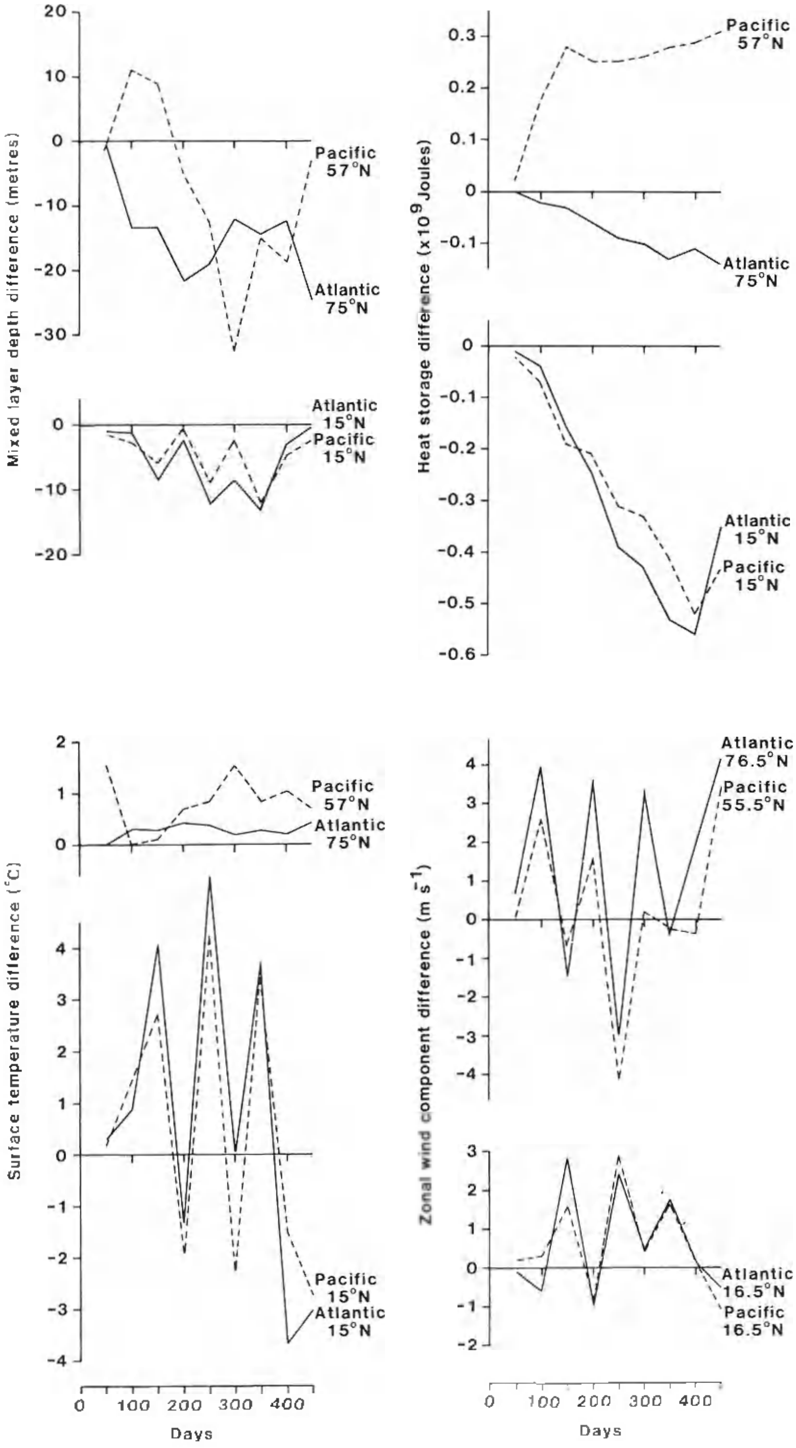


Fig. 3. Time series of 50-day averages from experiments A1 and A2 for mixed-layer depth and heat content of the mixed layer at latitudes in the northern and southern regions of each ocean. The series are averaged zonally across each ocean and expressed as deviations from the corresponding series of the control run.



Mixed layer depth difference (metres)
Pacific 57°N
Atlantic 75°N
Atlantic 15°N
Pacific 15°N
Heat storage difference (x10^9 Joules)
Pacific 57°N
Atlantic 75°N
Atlantic 15°N
Pacific 15°N
Surface temperature difference (°C)
Pacific 57°N
Atlantic 75°N
Pacific 15°N
Atlantic 15°N
Zonal wind component difference (m s^-1)
Atlantic 76.5°N
Pacific 55.5°N
Atlantic 16.5°N
Pacific 16.5°N
Days
Days

$$\frac{\partial}{\partial t}\,[(T_* - T_h)h^2] \;=\; G - D + S/\gamma \tag{3}$$

Figures 3 and 4 indicate that the total heat content of the mixed layer does not oscillate but shows merely an almost linear variation with time. This implies that on

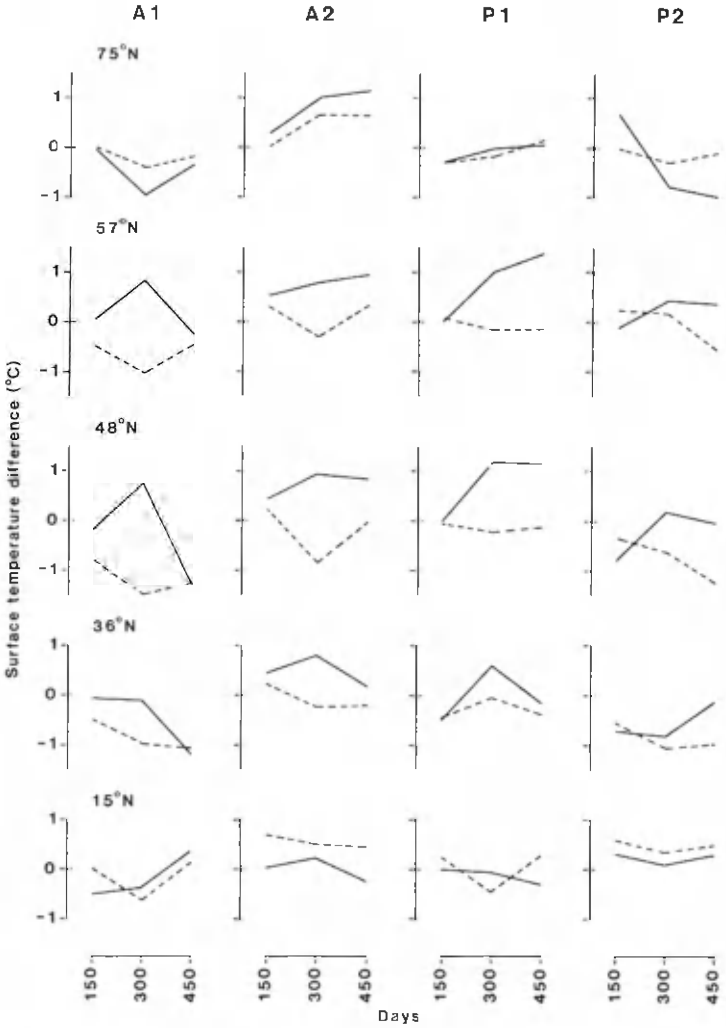


Fig. 5. Time series of 150-day averages from experiments A1, A2, P1 and P2 for surface temperatures at 15°, 36°, 48°, 57° and 75°N. The series are averaged zonally across each continent and expressed as deviations from the corresponding series for the control run. The variation of 50-day means about these values is ± 0.75° at mid-latitudes, lower at the highest latitudes and ± 0.25° at low latitudes.

Fig. 4. Time series of 50-day averages from experiment P1 for mixed-layer depth, heat content of the mixed layer, surface temperature and component of zonal wind at latitudes in the northern and southern regions of each ocean. The series are averaged zonally across each ocean and expressed as deviations from the corresponding series for the control run.

occasions the mixed layer can shallow without changing its heat content, which is an unrealistic feature of the model. However, even with a more accurate treatment of potential-energy losses, it is likely that fluctuations analagous to those shown here will occur. Thus, the thermal forcing is approximated as constant in time. In the model D was written as $D'h$ where D' was a decreasing function of latitude. Applying eqns. (2) and (3) to each ocean, a pair of equations describing the transient behaviour of the oceanic mixed layers in the model can be derived:

$$(S + B_A - M_A)h_A + (S + B_A - M_A)t\,\frac{\partial h_A}{\partial t} + H_A\,\frac{\partial h_A}{\partial t} = G_A - D'h_A + S/\gamma \qquad (4)$$

$$(S + B_P - M_P)h_P + (S + B_P - M_P)t\,\frac{\partial h_P}{\partial t} + H_P\,\frac{\partial h_P}{\partial t} = G_P - D'h_P + S/\gamma \qquad (5)$$

where suffix "A" refers to the Atlantic and "P" to the Pacific, and H_A and H_P are the initial heat contents of the two oceans' mixed layers. Equations (4) and (5) are coupled by the rates of wind mixing, G_A and G_P; each of these will be a complicated function of the temperature and mixed-layer depth distributions in the two oceans. It is these functions which will determine the periods of any oscillations. The dissipation terms $D'h_A$ and $D'h_P$ will tend to dampen any transients, and this effect will decline northwards with D'. As a result both the trend and the oscillation are most pronounced at high latitudes (Figs. 3 and 4).

An oscillatory teleconnection is also present in the continental-surface temperatures. These temperatures are much more variable in time than oceanic temperatures, so that the effect is most clearly illustrated by using 150-day averages (Fig. 5). While temperature fluctuations at the highest and lowest latitudes are similar over the two continents, in middle latitudes (45°–60°N) America and Eurasia show variations that are out of phase. In general, the temperature over each continent tends to be similar to that of the ocean upstream of it (i.e. to the west in the westerly wind belt).

DISCUSSION

Changing the oceanic heat flux in the model leads to a transient response with a distinct structure which shows coupling between the oceans. Although at low latitudes little or no longitudinal variation is shown, in middle and high latitudes there is a pronounced zonal asymmetry, with an east–west seesaw of period about 400 days occurring in each of the four runs. The mixed layers of the northern regions of the ocean and their response to varying wind-mixing and varying heat storage form an important element of this oscillation. The oscillation is also reflected in continental-surface temperatures. Having only one cycle per experiment, we cannot conclude that this mode is truly periodic; it may be a temporary feature of the adjustment process. Van Loon and Rogers (1978) have described a seesaw between winter temperatures at Greenland and those in northern Europe. This pattern is related (Rogers and Van Loon, 1979) to the strength of the zonal winds and is accompanied by large anomalies in the atmosphere–ocean–ice system, some of which persist through the subsequent spring and summer. The idealised calculations presented here suggest that such a seesaw effect could be a transient mode of the climatic system, perhaps triggered by an alteration of the oceanic heat transport.

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