A modelling and observational study of the relationship between sea surface temperature in the north-west Atlantic and the atmospheric general circulation

By T. N. PALMER and SUN ZHAOBO

Meteorological Office, Bracknell

(Received 24 January 1985; revised 4 July 1985)

SUMMARY

Results from four pairs of 50-day wintertime integrations of the Meteorological Office's 5-level general circulation model, with warm and cold sea surface temperature (s.s.t.) anomalies of about 3 K in the north-west Atlantic, are described. Difference fields between the warm and cold integrations are statistically significant at the 1% level with positive geopotential height over the central north Atlantic, and weaker negative height over Europe. The storm track over the Atlantic is displaced from its normal position. Results from four further pairs of integrations with halved s.s.t. anomalies are also described. The response is approximately linear, with systematic differences in 500 mb geopotential height over the Atlantic, parts of which are just significant at the 10% level with half the full s.s.t. anomaly. Overall, however, the model's response is weaker than could be obtained with tropical s.s.t. anomalies of the same magnitude.

Results from the model integrations are compared with results from an observational study of the relationship between wintertime s.s.t. in the north-west Atlantic, and mean sea level pressure and 500 mb height. Two independent 30-year periods were chosen for study, thus minimizing the influence of long-term trends in s.s.t. Over the Atlantic and Europe the model results compared well with the observations. With s.s.t. data lagging the atmosphere by one month, the observational study appears to show that the s.s.t. anomalies are initially forced by perturbations in the atmospheric circulation. With s.s.t. data leading the atmosphere by one month results show that atmospheric and s.s.t. anomalies are most persistent in the period October to December. Throughout the winter these lagged relationships are much weaker and not statistically significant.

Diagnostics of E-vector divergence from the GCM experiments are used to suggest that anomalous baroclinic wave activity over the Atlantic is important as a momentum forcing for the anomalous time-mean flow pattern. On the other hand, the role of thermal forcing, provided by anomalous diabatic heating and transient eddy heat flux convergence, may be important. To substantiate this statement, a simple linear steady-state two-layer model of the response to extratropical thermal forcing is described. With a suitable basic state, and a mid tropospheric heat source (given mainly by the transient eddy heat flux convergence), the response is shown to be equivalent barotropic with a downstream ridge and ascent over the thermal source.

Conversely, results from an ocean mixed layer model are discussed which suggest that warm s.s.t. anomalies could be maintained by a positive surface pressure perturbation positioned downstream of the anomaly, through anomalous northward advection of warm ocean water by Ekman drift currents. This northward advection would balance the sensible and latent heat loss into the atmosphere over the s.s.t. anomaly. Hence it is possible that some positive ocean-atmosphere feedback may account for the persistence of such atmospheric and oceanic anomalies.

1. INTRODUCTION

It is now well established, both from observations and modelling studies, that an anomalous warming of the surface waters of the tropical oceans, such as occurs during an El Niño event, can have a significant influence on the general circulation, both in the tropics and in the extratropics (see, for example, Horel and Wallace 1981; Blackmon et al. 1983; Palmer and Mansfield 1985a, b). On the other hand, the influence on the atmosphere of mid-latitude sea surface temperature (s.s.t.) anomalies is not well established. In the North Pacific, for example, whilst Namias (1973) has found significant correlations between atmospheric fluctuations and underlying s.s.t. variability, general circulation model (GCM) studies have failed to find a significant atmospheric response to a realistic North Pacific s.s.t. anomaly (see, for example, the recent study of Pitcher et al. (1985); but see also a critique of this paper by Namias and Roads (1985)).

1 Permanent address: Department of Meteorology, Nanjing Institute of Meteorology, Nanjing, The People's Republic of China
One of the regions where s.s.t. anomalies can be particularly large (up to 3 K in winter with larger values at other times of the year) is near the interface of the Gulf Stream and the Labrador Current, to the south-east of Newfoundland. One of the most active regions of the extratropical atmosphere, dynamically and thermodynamically, occurs above this region. In winter, with large air–sea temperature differences and strong meridional s.s.t. gradient (see Fig. 1), cyclogenesis is intense in this region (for example, according to Lau et al. (1981) band-pass eddy temperature and geopotential height variances in the mid troposphere are larger above this region than anywhere else in the northern hemisphere). Namias (1964) has found that blocking activity over northern Europe during the late 1950s was associated with abnormally cold water near the coast of Newfoundland. Namias (1973) has also associated the unusual conditions over the Atlantic and Europe in 1972 with warm (up to 6 K) s.s.t. anomalies off the eastern U.S. seaboard. Ratcliffe and Murray (1970) (hereafter RM) claimed to have found significant lagged correlations between s.s.t. in this area (hereafter the RM area) and surface pressure over the Atlantic and Europe, suggesting that a knowledge of s.s.t.s in the RM area may be of importance for long-range forecasting over Europe. Further observational evidence for RM’s conclusions was given by Folland et al. (1982) using multivariate statistical techniques, and s.s.t. in the RM area is currently used as one predictor in the statistical long-range forecasting model used in the Meteorological Office.

Corroborative evidence from GCM integrations with s.s.t. anomalies in the north-
west Atlantic is equivocal, however. Houghton et al. (1974) studied the effect of warm and cold anomalies of approximately 2 K in the RM area using an NCAR GCM. The model was run for 80 days starting from an isothermal initial state. Whilst they were able to find an individual 30-day period when the model’s mean surface pressure field compared well with data supplied by Ratcliffe and Murray, individual 30-day periods showed considerable variability. Overall, internal variability in the model was such that the results were not statistically significant. Kutzbach et al. (1977), using the same model, reported that a statistically significant signal was obtained over the Atlantic at 6 km with a warm s.s.t. anomaly twice the magnitude of that used by Houghton et al. This signal comprised a positive 6 km pressure anomaly over the central North Atlantic. Overall, however, multiple integrations from different initial conditions are required to isolate possible effects of such s.s.t. anomalies in a GCM, with confidence.

There are also a number of reasons why further assessment of the observational evidence may be desirable. Firstly, in the decade and a half since RM’s work, the Meteorological Office historical s.s.t. dataset (MOHSST) has been vastly improved (Minhinick and Folland 1984). Secondly, the interpretation of RM’s results may have been obscured by a trend in global observations of s.s.t. over the last century. Part of this trend results from changes in the method of measuring s.s.t.; part of it may also represent a real change in global s.s.t. (Folland et al. 1984). Using data quoted in Houghton et al. (1974), 10 of RM’s 13 warm anomaly cases occurred after 1949, and all of their nine cold cases occurred before 1921. According to Folland et al. globally averaged s.s.t. increased from the beginning of the century to a maximum in the 1940s, with a weak decline thereafter. It is possible that this trend may have influenced RM’s choice of warm and cold cases. Thirdly, RM published only the results of associations between s.s.t. and surface pressure, with surface pressure lagging s.s.t.s by one month. In order to help clarify the causal relationships between ocean and atmosphere, it is of interest to examine both synchronous and lagged associations of the opposite sense. Furthermore it will be useful to extend the atmospheric diagnostics to upper air observations where the magnitude of atmospheric anomalies may be larger.

In view of this, and of the potential importance for long-range forecasting of establishing the relationship between north-west Atlantic mid-latitude s.s.t. anomalies and the atmospheric general circulation, it was decided to conduct a further investigation into the model and observational results.

In order to establish whether a clear signal could be obtained from a model experiment, four pairs of 50-day integrations were run on the Meteorological Office’s 5-level GCM (Corby et al. 1977) with warm and cold s.s.t. anomalies whose magnitudes were about half as large again as the composite warm winter anomalies reported by RM. Results are presented as differences between warm and cold anomaly integrations. Unlike the integrations in Houghton et al. (1974), the model was not initialized with an isothermal atmosphere, but with observed analyses from four different years. Four further pairs of integrations were performed with halved s.s.t. anomalies.

The observational study used the MOHSST dataset together with daily historical grid point p.m.s.l. (pressure at mean sea level) and 500 mb analyses held at the Meteorological Office (Parker 1980). In order to minimize the global trends in s.s.t. observations mentioned above, independent estimates of the relations between s.s.t. and atmospheric circulation were made from data for 1951–1980 and 1901–1930. Synchronous and time-lagged relationships were investigated.

Results from both the observational and modelling study indicate that a statistically significant response in the atmospheric circulation can be associated with s.s.t. anomalies in the RM area, not only over the Atlantic, but further downstream over Europe.
Furthermore the GCM response is in general agreement with the appropriate observations.

Results from the asynchronous observational study, and a mixed layer ocean model, suggest that the s.s.t. anomalies are initially forced by atmospheric circulation perturbations, probably through the effects of anomalous sensible heating of the mixed layer and advection of mean s.s.t. by anomalous Ekman drift currents. On the other hand diagnostics from the GCM and results from a simple two-level model of the atmosphere indicate that these atmospheric anomaly patterns can themselves be partially maintained through the ocean's influence on diabatic heating in the atmosphere, and on the position and intensity of the growth of baroclinic waves in mid-latitudes. Hence there appears to be the possibility of air/sea interaction feedback which can help the anomalies develop and persist. Observations suggest that this feedback process is particularly effective during November and December.

Overall it is found that the influence on the atmospheric general circulation of s.s.t.s in the RM area is certainly not as strong as can be obtained with anomalies of similar magnitude in the tropics (see Palmer and Mansfield 1985a, b), and, for example, the effects of warm and cold anomalies whose magnitudes do not exceed about 2 K will, in practice, be too weak to detect above the internal variability of the atmosphere.

In sections 2 and 3 we describe the GCM and observational results respectively. In section 4 these results are compared, and in section 5 some theoretical discussion is given to try to understand both how the ocean may be influencing the atmosphere, and vice versa.

2. Model results

In this section we first describe the results from eight 50-day wintertime integrations on the hemispheric version of the Meteorological Office 5-level model (Corby et al. 1977). The model has a uniformly spaced horizontal grid with a resolution of 330 km, and was run with fixed zonally symmetric wintertime radiation constants. The integrations were initialized from November analyses for four consecutive years (11 November 1978, 7 November 1979, 21 November 1980, 20 November 1981). Our choice of initialization data was restricted as only a limited number of suitable analyses had been archived. For each initialization date two integrations were run. One had the (positive) s.s.t. anomaly illustrated in Fig. 1(a) added to climatological wintertime s.s.t.s, the other had this anomaly subtracted from the climatological s.s.t.s. This idealized s.s.t. anomaly was based on the composite warm anomaly for a number of December observations, as reported by RM (see also Fig. 17 below). In RM's composite the largest anomalies have magnitudes in excess of 2 K. Recognizing the difficulty of previous workers to find a significant response to mid-latitude anomalies of 1 or 2 K it was decided at the outset to impose an enhanced version of the RM December composite into the model. The atmospheric response to weaker anomalies will be discussed below. In Fig. 1(a) there is a small area in excess of 3 K, with a maximum of 4 K representing an enhancement of about 1.5 over the RM December warm composite. The position of the centre of the anomaly occurs in the region of maximum climatological meridional s.s.t. gradient. The effect of adding (subtracting) the s.s.t. anomaly to (from) the climatological s.s.t.s is shown in Fig. 1(b), (c).

In order to avoid consideration of the artificial initial transient adjustment of the model to the imposed s.s.t.s, we present difference fields (positive anomaly run minus negative anomaly run) for the last thirty days of each fifty-day integration. By looking at these difference fields, rather than the difference between the positive anomaly runs
Figure 2. (a) 500 mb geopotential height difference field (dam), averaged over the four pairs of integrations. Each pair of integrations consists of one 50-day run with the s.s.t. anomaly shown in Fig. 1, and one run with the negative of this s.s.t. anomaly. For this and other fields the mean of days 21-50 are shown. (b) Student $t$ values for the mean difference field. (c) Mean 500 mb geopotential height (dam), averaged over all eight integrations.

and a control integration with climatological s.s.t.s, the atmospheric response should be further amplified against the model's internal variability. A similar procedure has been used for the observational study.

(a) The stationary response

The 500 mb (30-day) time-mean geopotential height difference field, averaged over the four pairs of experiments, is illustrated in Fig. 2(a). Over the Atlantic there is a positive centre with a magnitude of 12 dam at 30°W 47°N. Further downstream, over central Eastern Europe there is a negative centre with a magnitude of 5 dam. A third positive centre with magnitude 4 dam lies close to Lake Balkhash in the U.S.S.R.

Since each pair of integrations is independent of all other pairs (initialization dates were one year apart), a $t$-test can be performed on the integrations to assess the statistical
significance of this difference field. This is shown in Fig. 2(b). With a sample size of eight integrations there are six degrees of freedom in the $t$-test. The hypothesis that the difference fields are not significantly different from zero will be rejected at the 1% confidence level if $t > 3.7$; at the 5% level if $t > 2.4$; and at the 10% level if $t > 1.9$ (using two-sided $t$ values). It can be seen that the positive height difference over the Atlantic is highly significant, and much of it is contained within the 5% contour. A substantial part
of the difference field is also significant at the 1% level. It is interesting to note that the maximum value of $t$ is positioned some 30 degrees downstream of the s.s.t. anomaly and is 10 degrees east of the positive centre over the Atlantic, signifying that at 45°N the model's internal variability is smaller over the east than over the central Atlantic.

The negative height difference values over Europe are less significant than those over the Atlantic, with only a small area significant at 5%. Most of the negative anomaly, however, is significant at 10%. The positive centre over the U.S.S.R. appears to be significant at the 10% level, though the significance apparently increases to the south.

In Fig. 2(c) we show the mean 500 mb flow averaged over all eight integrations. Over the central Atlantic the meridional gradient is about 30% stronger than observation (see Lau et al. 1981) implying excessively westerly zonal winds. This is likely to influence the wavelength of the response in Fig. 2(a) (see section 4).

In Figs. 2(d)–(g) we show the 500 mb geopotential height difference field for each of the four pairs of integrations. These diagrams essentially confirm the results of the $t$ statistic: for each pair of integrations positive values are clearly visible over the central Atlantic, and negative values over Europe. The positive centre varies little in position between pairs of runs, though the position of the negative centre is more variable.

For brevity, we have not illustrated the differences in wind, though it is clear from Fig. 2 that associated with the warm s.s.t. anomaly there are enhanced westerlies in the mid Atlantic to the north-east of the s.s.t. anomaly and reduced westerlies in the mid Atlantic to the south-east of the s.s.t. anomaly. This can be interpreted in terms of a tendency to shift the climatological jet stream to the north of the s.s.t. anomaly. Over the RM area the anomalous geostrophic winds are south-easterly, over the British Isles they are north to north-westerly.

The time-mean 500 mb temperature difference field, averaged over the last thirty days of the four pairs of experiments, is illustrated in Fig. 3(a). Three temperature

![Figure 3](image-url)
anomaly centres are associated with the geopotential height centres discussed above. The strongest centre, over the Atlantic, is positioned about ten degrees to the west of the height centre. As with the height anomalies, the positive centre over the Atlantic is significant at the 1% level, whereas the negative centre over Europe is significant only at the 10% level. The mean 500 mb temperature field for all eight integrations is illustrated in Fig. 3(b). It can be seen from Figs. 2(a), 2(c), 3(a) and 3(b) that the advection of anomalous temperature by the mean wind is largely cancelled by the advection of mean temperature by the anomalous wind.

The time-averaged difference in precipitation for the four pairs of experiments is illustrated in Figs. 4(a)–(d), with largest values (about 10 mm d\(^{-1}\)) approximately over the s.s.t. anomaly. Surface evaporation of about equal magnitude occurs in the same region. It is likely that a large contribution to this time-averaged precipitation anomaly may occur during cold outbreaks associated with the development of mid-latitude cyclones over the anomalously warm ocean. In other words, the time-mean precipitation anomaly may be strongly influenced by the strength of transient eddy activity over the s.s.t. anomaly (see also the next sub-section on eddy activity). Associated with this rainfall there is enhanced low-level convergence indicating ascent over the s.s.t. anomaly. It would appear likely in these experiments with a synoptic-scale s.s.t. anomaly, that adiabatic cooling by ascent over the s.s.t. anomaly may play a major role in balancing the effects of latent heat release (see the discussion in section 5).

![Figure 4](image-url) Rainfall difference fields (mm d\(^{-1}\)), for each of the four pairs of integrations.
The 1000, 850 and 250 mb mean geopotential height difference fields are plotted in Figs. 5(a)–(c). At 1000 mb, the positive Atlantic centre has a magnitude of about 8 dam, corresponding to a p.m.s.l anomaly of about 10 mb. This centre is positioned at about 15°W, which is 15° to the east of the corresponding centre at 500 mb. Most of this westward tilt with height appears to be confined to the model's boundary layer; at 850 mb this positive anomaly centre is located close to 25°W. Between 500 and 250 mb the difference field is very closely equivalent barotropic.

At 1000 mb, over the s.s.t. anomaly, there is a separate negative centre, significant at the 5% level, which is only weakly apparent at 850 mb, and absent at 500 mb. Associated with this low-level feature is a positive temperature anomaly in the boundary layer (not shown). It would appear that this shallow feature arises from a warming of the air above the s.s.t. anomaly as a result of the boundary-layer physics and anomalous latent heating (see section 5).

(b) Transient eddy fields

We have already considered the possibility that the passage of developing mid-latitude depressions over the s.s.t. anomaly may account for a substantial part of the

Figure 5. Geopotential height difference (dam), averaged over the four pairs of integrations. (a) At 1000 mb; (b) at 850 mb; (c) at 250 mb.
precipitation enhancement illustrated in Fig. 4. Another possible influence of the baroclinic waves is through their feedback by heat and momentum flux convergences onto the time-averaged flow.

In view of these considerations, filtered values of daily variance of 500mb temperature, geopotential height and $E$-vector divergence (see below) were calculated. The filtering is achieved by computing the variance of daily fields from non-overlapping 5-day periods, and averaging these over the six 5-day periods comprising days 21–50 of each experiment. This method of filtering has been described by Lorenz (1979) as the 'poor man's filter' and has been used recently by Hoskins et al. (1983). The filter emphasizes fluctuations with time scales appropriate to baroclinic wave activity.

In Fig. 6, values of mean difference of temperature variance between warm and cold integrations are plotted. In the mid Atlantic, over and to the north of the s.s.t. anomaly, the variance of temperature is increased (relative to the cold s.s.t. anomaly) by up to 12 K$^2$. The maximum increase occurs in the region of enhanced mean westerlies (see Fig. 2(a)). In the mid Atlantic, to the east of the s.s.t. anomaly, the temperature variance is

![Figure 6](image_url)

Figure 6. (a) Difference field of transient eddy temperature variance (K$^2$) at 500 mb, filtered to remove low-frequency variability, averaged over the four pairs of integrations. (b) Mean eddy temperature variance (K$^2$) at 500 mb, filtered to remove low-frequency variability, averaged over all eight integrations.
decreased by up to 2 K$^2$, coincident with the region of anomalous mean easterlies. Over the British Isles and Europe, the anomalous temperature variance is positive. Mean temperature variances, averaged over the eight experiments, are plotted in Fig. 6(b), which reflects the normal position of the Atlantic storm track. The maximum values in the Atlantic storm track are 12 K$^2$, and the position and magnitude of these values correspond well with the observed winter climatology reported by Lau et al. (1981). (As mentioned in the introduction, it is interesting to note that in Lau et al.'s data, the largest northern hemisphere values of band-pass 500 and 850 mb temperature variance occur directly over the RM area.) The anomalies in Fig. 6(a) therefore represent sizeable departures from these average values.

The anomalous filtered geopotential height variance is not illustrated. As with the temperature variance, values are generally positive to the north of the s.s.t. anomaly, and continue to be positive downstream over the British Isles and Europe. In the mid Atlantic, to the south-east of the s.s.t. anomaly, values are negative.

A final eddy statistic to be presented is the filtered horizontal $E$-vector divergence difference field at 500 mb (Fig. 7). The horizontal $E$-vector (not strictly a vector) is defined by

$$
E = (\overrightarrow{u'^2} - \overrightarrow{u^2}, -\overrightarrow{u'v'})
$$

where (for this equation only) overbars denote time averages and primes departures therefrom. Whilst the full $E$-vector has a vertical component, representing effects from baroclinic processes, the divergence of the horizontal component of the $E$-vector provides a qualitative measure of the depth-averaged eddy forcing on the time-averaged wind (Hoskins et al. 1983). Figure 7(b), which shows the mean $E$-vector divergence over the mean warm and cold integrations, shows a band of positive values with magnitude up to $6 \times 10^{-5}$ m$^{-2}$ extending north-west from Newfoundland across the Atlantic to Iceland. This corresponds to the GCM storm track across the Atlantic and signifies the well-known tendency of eddies to accelerate the mean flow in their growing phase. With the mean jet of about 25 m s$^{-1}$, the eddy forcing time scale on the mean flow ($-\nabla \cdot \overrightarrow{E}$) of 4–5 days is probably sufficient to offset the effects of dissipation on the mean flow.

Figure 7(a) shows the filtered $E$-vector divergence difference field (warm–cold cases). Near the s.s.t. anomaly is a dipole with magnitudes comparable with the mean $E$-vector divergence field shown in Fig. 7(b). Over and to the east of the RM area is an area of convergence with magnitude up to $-6 \times 10^{-5}$ m$^{-2}$. To the north and north-east of the RM area are areas of divergence with magnitudes up to $4 \times 10^{-4}$ m$^{-2}$. These values imply an eddy-induced forcing on the anomalous time-mean flow with a time scale ($\delta v/\delta \nabla \cdot \overrightarrow{E}$) of only a couple of days, which appears to imply that the forcing is very substantial. (Here $\overrightarrow{\delta X}$ denotes the difference in $X$ between the warm and cold integrations.) These difference fields are not highly significant, though the area where the anomalous convergence exceeds $5 \times 10^{-5}$ m$^{-2}$ near the RM area is significant at 5%. The area of anomalous divergence to the north and north-east of the RM area appears less significant.

It is possible that enhanced gradients to the north of the s.s.t. anomaly gave rise to anomalous baroclinic development, this in turn forcing stronger mean westerlies, though it is impossible to argue a causal relationship from the GCM diagnostics. All that can be established is that the weaker (stronger) mean westerlies to the south- (north-) east of RM area are associated with a weakening (strengthening) of high-pass eddy temperature and geopotential height variance, and a decrease (increase) in the nonlinear mean flow forcing by the eddies.
The effect of anomalous transient eddy heat fluxes (the vertical component of the $E$-vector) on the mean flow is considered in section 5.

(c) Integrations with reduced s.s.t. anomalies

In order to obtain some estimate of the dependence of the atmospheric response on the magnitude of the s.s.t. anomaly, four further pairs of 50-day integrations have been run with plus and minus half the anomaly shown in Fig. 1. The initial data for these integrations are the same as those with the full s.s.t. anomaly.

The rainfall difference fields (not illustrated) show an almost linear response to the s.s.t. anomaly. The 500mb difference field also has an approximate linear response. Figure 8(a) shows this difference field for the half-s.s.t. anomaly experiments averaged together, and Fig. 8(b) shows $t$ values associated with this difference field.

Over the mid Atlantic there is a positive height anomaly with a magnitude of 6 dam (compared with 12 dam for the full s.s.t. anomaly), part of which is just significant at the
10% confidence level. Over eastern Europe there are weak negative height anomalies which possibly are just significant.

Hence, a linear scaling of the response with the magnitude of the s.s.t. anomaly may not be grossly inaccurate. It should be remarked that such a scaling implies that an s.s.t. anomaly with an area-averaged value of only 1 K would be associated with a 500 mb height anomaly of only 2 dam (compared with climatology), a value which would not be readily discernible above internal variability.

3. OBSERVATIONAL RESULTS

As discussed in the introduction, there are a number of reasons why it is desirable to conduct a further assessment of the observational relationship between the atmospheric general circulation, and s.s.t.s in the RM area. An account of this study is given below.

Sea surface temperature data were obtained from the monthly mean Meteorological Office historical s.s.t. dataset, and atmospheric 500 mb and p.m.s.l. fields were obtained from historical daily northern hemisphere analyses kept at the Meteorological Office. Two different 30-year periods have been chosen for this study, 1951–1980 and 1901–1930, and we have analysed the results from each period independently.

(a) Synchronous associations

We define the ‘RM area’ to be that contained within longitudes 60 and 40°W, and latitudes 50 and 40°N. For each month from November to February inclusive we have calculated normal, or climatological, global s.s.t.s for the 1951–1980 data, and, separately, for the 1901–1930 data. Relative to the appropriate monthly normal, we have calculated
s.s.t. anomalies for $10^6 \times 10^6$ squares over the globe, for every month from November to February inclusive, of every year within these two periods.

Monthly-mean s.s.t. anomalies greater than 0.8K, averaged over the RM area, are grouped together to form a composite warm s.s.t. anomaly. Similarly, monthly-mean s.s.t. anomalies less than -0.8K, averaged over the RM area, are grouped together to form a composite cold s.s.t. The ‘warm’ and ‘cold’ months thus chosen are tabulated in Tables 1 and 2. The difference between the warm and cold composite s.s.t. anomalies are illustrated in Fig. 9(a) for 1951–1980, and in Fig. 9(b) for 1901–1930 (blank squares indicate missing data).

The values shown in Fig. 9 clearly illustrate the effect of this compositing, with the $10^6 \times 10^6$ difference field having a maximum of 2.4K in the RM area. This value is consistent with the magnitude of warm and cold composite anomalies used by RM if these latter values are averaged over $10^6 \times 10^6$ squares. Also noticeable is the lack of a
significant anomaly (i.e. with magnitude greater than 1 K) in any other area, the tropics in particular.

Climatological values of monthly-mean 500 mb geopotential height and p.m.s.l. were calculated for the 1951-80 data, and p.m.s.l. for the 1901-30 data (500 mb data were not available for this period). For each month designated either 'warm' (s.s.t. anomaly in the RM area greater than 0·8 K) or 'cold' (s.s.t. anomaly less than −0·8 K) the 500 mb height and p.m.s.l. monthly mean anomalies were calculated. These were composited together to create mean synchronous atmospheric anomalies associated with the warm and cold composite s.s.t. anomalies.

The 500 mb geopotential height field difference between the composited warm and cold months for the 1951–1980 period is illustrated in Fig. 10. To the north-east of the RM area there is a positive centre with a maximum of 11 dam at 50°N 40°W. Over the coast of Norway there is a negative centre with a magnitude of −8 dam. A second positive centre with maximum of 14 dam is located over Siberia, and there is a second negative centre over the Pacific near the dateline. A t test has been performed on these fields treating each month that comprises the composite as an independent element. (This may give an overestimate of the significance, as some of the months are adjacent in time, see Table 1.) The areas enclosed by the 5% significance contour are shown stippled. It is seen that the centres over the Atlantic, northern Europe and Siberia are significant, though the low centre over the Pacific is not.
Figure 10. Observed composite 500 mb geopotential height difference field (dam), synchronous with the 'warm' and 'cold' months in Table 1, using monthly-mean data from 1951–80. Stippling indicates areas that are significant at the 5% confidence level.

Figure 11. As Fig. 10 for p.m.s.l. data (mb) from 1951–80.
The mean p.m.s.l. difference field for the warm minus cold months in the 1951–80 period is illustrated in Fig. 11. It can be seen that between the surface and 500 mb the response is approximately equivalent barotropic, with positive and negative pressure centres with magnitudes from 6 to 9 mb, positioned in approximately the same positions as at 500 mb. As with the 500 mb height field, the Atlantic, European and Siberian anomalies are significant at the 5% level.

The mean p.m.s.l. difference for the warm minus cold months from the independent period 1901–1930 is illustrated in Fig. 12. A positive centre (7 mb) over the Atlantic and a negative centre (6 mb) over Europe, in approximately the same positions as the p.m.s.l. difference fields for the 1951–1980, are readily apparent. These are both significant at the 5% level. The most obvious difference in the p.m.s.l. fields between the two periods is the absence of a strong positive centre over Siberia in the 1901–1930 data, though there is a weaker positive centre at about 105°E 45°N, which apparently is significant.

Putting the results from the two independent periods together, it appears that the positive centre over the Atlantic and the negative centre over northern Europe are highly significant, while the positive centre over Siberia is less so.

We have calculated the mean difference between the warm and cold months of the filtered 500 mb geopotential height variance but for lack of space this is not illustrated. The definition of the filter is exactly the same as was used to study eddies in the GCM integrations, and the main feature is the decrease in high-pass eddy variance to the east of the RM area, the anomaly centre having a magnitude of $-50 \text{dam}^2$, significant at the 1% level. There is some indication of enhanced eddy variances to the north-east of the RM area over Greenland, but these are less significant.
(b) Asynchronous associations

In order to try to estimate from the observations whether the s.s.t. anomalies are initially forced by the atmosphere or vice versa, we have composited together atmospheric anomalies the month before and the month after the occurrence of the warm and cold months used in section 3(a).

Specifically, the monthly-mean 500 mb height and p.m.s.l. anomalies were calculated for the month preceding each warm month in Tables 1 and 2. These anomalies were then composited together for 1951–1980 and separately for 1901–1930. A similar composite was produced for monthly-mean atmospheric anomalies following each warm month. The procedure was repeated for the cold months in Tables 1 and 2.

The 500 mb height difference field with the atmosphere preceding the warm and cold months is illustrated in Fig. 13, for the 1951–80 data. It can be seen that there is very little difference between this field and that shown in Fig. 10 for the synchronous relationship. In particular there is a highly significant positive anomaly with magnitude 12 dam over the Atlantic to the east of the RM area. The negative anomaly centres over Iceland and the subtropical Atlantic also appear to be significant. A similar pattern appears for the p.m.s.l. field a month preceding the warm and cold RM months for both 1951–80 and 1901–30 p.m.s.l. data. It appears, therefore, that the s.s.t. anomalies initially develop in response to anomalous atmospheric forcing. Figures 14(a), (b) show the s.s.t. difference map for the months preceding and following the composite in Fig. 9. In both diagrams there are positive differences in the RM area, these being somewhat larger for the preceding month than for the following month. In particular, Figs. 9 and 14 show that, on average, s.s.t. anomalies in the RM area do tend to be persistent over several months, at least in the early/mid winter period.

Figure 13. Observed composite 500 mb geopotential height difference field (dam) leading the 'warm' and 'cold' months shown in Table 1, using monthly-mean data from 1951–1980. Stippling indicates areas that are significant at the 5% confidence level.
A different pattern emerges for atmospheric anomaly patterns occurring a month after the warm and cold RM months. Figure 15 at 500 mb for 1951–1980 shows a weak low anomaly over the Atlantic, and a second over Europe. None of this pattern, however, is statistically significant. A similar pattern occurs for p.m.s.l. For the 1901–30 data there is a small positive p.m.s.l. anomaly over the Atlantic, but again this is not statistically significant.

Before commenting on this result we show in Figs. 16 and 17 a reproduction of two of RM’s figures, showing the (lagged) association between warm s.s.t. anomalies in the RM area and surface pressure anomalies one month later. Figure 16 is for the autumn period October/November, and Fig. 17 is for the winter period December/January. The stippling shows areas significant at the 5% level.

In Fig. 16 there are two significant centres which have clear counterparts in Figs. 11 and 12. On the other hand the lagged associations shown in Fig. 17 are much less significant. This suggests that atmospheric circulation anomalies, associated with s.s.t. in

![Figure 14](image-url) Observed composite s.s.t. differences ('warm' – 'cold') in units of one tenth K. Blank sea squares indicate missing data. (a) One month preceding the 'warm' and 'cold' months shown in Table 1. (b) One month following the 'warm' and 'cold' months shown in Table 1.
Figure 15. Observed composite 500 mb geopotential height difference field (dam) lagging the 'warm' and 'cold' months shown in Table 1, using monthly-mean data from 1951–1980.

the RM area, are more likely to be persistent in autumn and early winter than in mid and late winter.

This conclusion is consistent with the findings of Folland et al. (1982), and the observational results in this paper. As shown in Tables 1 and 2, most of the warm and cold months from the composite occur in November and December rather than January and February. More generally, during the winter months, the standard deviation of interannual variability of monthly-mean s.s.t. in the RM area is largest in November and decreases to a minimum in February (M. R. Newman, personal communication). Furthermore, according to Tables 1 and 2, there appears to be less persistence of large anomalies from December to January, as compared with November to December. For example, none of the 'warm Januaries' shown in Table 1 for 1951–1980 were preceded by 'warm Decembers'. On the other hand, three 'warm Decembers' (1951, 1953 and 1967) succeeded 'warm Novembers', two of which (1951, 1967) were, in fact, preceded by 'warm Octobers'.

How true these results are in general requires further investigation; however, it appears that large s.s.t. anomalies in the RM area are more likely to occur during late autumn and early winter, than during mid to late winter. A possible reason for this is that large s.s.t. anomalies are difficult to achieve in late winter when the ocean mixed layer is deepest.

In conclusion, therefore, the lagged associations from both RM and this paper appear to indicate that the anomalous atmospheric circulations are important in forcing s.s.t. anomalies in the RM area, and that these circulations are persistent in the autumn and early winter period (October–December), though not in the later winter period (January–February).
4. COMPARISON BETWEEN OBSERVATIONS AND MODELLING RESULTS

In this section we shall compare results from the last two sections, and from other modelling studies.

In comparing the 500 mb response between the model (with full s.s.t. anomaly) and observational studies in this paper, we note that the model has a maximum of 12 dam positioned at 47°N 30°W (Fig. 2), whereas the observational counterpart has a maximum of 11 dam positioned at 50°N 42°W (Fig. 10). The existence of a strong downstream high extending over much of the North Atlantic in both studies is, of course, extremely encouraging. Whilst the model's response is similar to the observations, one must allow for the fact that the model s.s.t. anomaly has a magnitude about twice as large as the observed composite s.s.t. anomaly. (The runs with the more realistic halved s.s.t. anomaly show a maximum of about 6 dam.) There is a discrepancy of about 10° of longitude in the position of the centres in Figs. 2 and 10, though over the RM area the 500 mb wind anomalies for both model and observations have a southerly component. There are two possible reasons for this. Firstly, the discrepancy in position may be due to the possibility
in the atmosphere of additional persistent thermal forcing over north-east Canada associated with the early onset of snow cover in years with cold s.s.t. anomalies or the late onset of snow cover in years with warm s.s.t. anomalies. In the model, with fixed winter radiation conditions, the ground over north-east Canada is frozen for both warm and cold integrations. Secondly, the upper-level flow over the North Atlantic is, in common with many GCMs, too strong over the Atlantic. Thus the stationary Rossby wavelength in the model will be too long, and this may affect the position of the model’s downstream response. Studies with tropical s.s.t. anomalies (Palmer and Mansfield 1985b) have shown this to be important.

Both model and observations show a negative centre over Europe, though the model’s response is both weaker and positioned further south than in the observations. Nevertheless, both show a north-westerly wind anomaly over the U.K. with a height difference across the British Isles of 5 dam for the model and 4 dam for the observations. Over the U.S.S.R. the model and observations are in poor agreement, though as Figs. 11 and 12 show, the surface pressure anomalies between the two 30-year periods are not in good agreement over the U.S.S.R; also as mentioned in Corby et al. (1977), the model’s climatology is poor over central Asia.
The surface pressure anomalies over the Atlantic for the two 30-year periods agree well with each other. Both show maxima of about 7 mb at about 50°N and 35-40°W. Again the model shows an anomaly maximum which is slightly larger than the observations (8 dam ~ 10 mb), and positioned about 20° too far to the east. As discussed in section 2 there is a phase shift with height in the model boundary layer (~10° between 1000 and 850 mb) which is not apparent in the observations. Whether this discrepancy arises for numerical reasons, as a result, for example, of the model’s coarse vertical resolution, or inadequacies in the boundary layer physics is not known. However, as with the 500 mb height anomalies, the surface pressure gradient over the U.K. is such as to give a north to north-westerly surface wind anomaly with a 4 mb difference over the British Isles for both model and observations.

Near the s.s.t. anomaly the model results show a weak negative height anomaly at 1000 mb which is contained within the model boundary layer. There is no obvious counterpart to this in the surface pressure observations shown in Figs. 11 and 12. (The only possible indication of this effect in the observations is the slight cyclonic curvature of the anomalous surface isobars over the RM area in Fig. 12, opposite to the curvature of the anomalous 500 mb height contours over the RM area in Fig. 10.) It is possible that this negative height anomaly which depends on the boundary layer parametrization scheme (see section 5) has no counterpart in the real atmosphere. On the other hand, bearing in mind the area this anomaly covers, it is also possible that the observational network would be unable to resolve such a feature with any degree of reliability.


Bearing in mind the model's crude boundary layer parametrizations (see Corby et al. (1977) and the next section), its coarse vertical resolution, and its poor climatology over Asia, the model and observational study appear to be in satisfactory agreement.

Finally, in this section, we shall make a few remarks about how our results compare with Houghton et al. (1974), Kutzbach et al. (1977) and Pitcher et al.'s (1985) studies of the response in a GCM to extratropical s.s.t. anomalies.

As mentioned above, Houghton et al. studied the response to an s.s.t. anomaly in the RM area with maximum amplitude of 2 K. When comparing with a control integration they appeared not to be able to find a statistically significant response in surface pressure. There are two principal reasons why our results are not directly comparable with those of Houghton et al. Firstly, our integrations were longer (four pairs of 50-day runs as opposed to one pair of 80-day runs), and were initialized with real data, as opposed to isothermal conditions. Secondly, we used an enhanced s.s.t. anomaly, and compared warm and cold integrations. On the other hand, Kutzbach et al. (1977) have reported a statistically significant response in the NCAR GCM at 6 km, with an enhanced version of Houghton et al.'s warm s.s.t. anomaly. Over the central North Atlantic, Kutzbach et al. report anomalous positive pressure, in general agreement with our results.

Pitcher et al. (1985) studied the response of the NCAR spectral GCM to a mid North Pacific s.s.t. anomaly (taken from observations from the winter 1976/7), with 1200-day perpetual January simulations. With an enhanced s.s.t. anomaly they find a weak response which bears some similarity to the Pacific/North American teleconnection pattern (Wallace and Gutzler 1981). Insofar as the 700 mb response to the enhanced cold anomaly comprised a negative height anomaly downstream, with north-easterly winds over the s.s.t. anomaly, there is some agreement with the present results. Namias and Roads (1985) have argued that the effect of mid-latitude s.s.t. anomalies in Pitcher et al.'s results were at least as important as tropical El Niño s.s.t.s for specifying the observed anomalous atmospheric features of the 1976/7 winter.

It is worth commenting that if the strength of transient eddy activity is fundamentally important in determining the time-mean response (see section 5), then GCMs whose
horizontal resolution is not sufficient to resolve adequately the effects of such transient eddies, may underestimate the atmospheric response to an extratropical s.s.t. anomaly.

5. Theoretical Considerations

The results of the model integrations in section 2 showed that the large-scale atmospheric circulation over the Atlantic can be influenced by sufficiently large s.s.t. anomalies in the RM area. The results of the synchronous and lagged relationships in section 3 indicated that anomalies in the atmospheric circulation could force changes in s.s.t. in the RM area both of which can persist through the early winter. A comparison of model and observational results in section 4 showed sufficiently good agreement that it appears possible that some form of air–sea interaction could help both atmospheric and ocean anomalies develop and persist over several months. In this section we discuss mechanisms that may account for these results.

(a) The influence of mid-latitude s.s.t. anomalies on the atmosphere

In Fig. 7 we showed diagnostics of E-vector divergence which suggested that momentum forcing by the transient eddies on the mean flow was a major component in accounting for the model's downstream response to the s.s.t. anomaly.

On the other hand it is possible that the response to thermal forcing is another component which cannot be neglected. It might well be thought, however, that an equivalent barotropic atmospheric anomaly pattern with a positive height downstream of the thermal forcing region is not consistent with thermal forcing theory (see Smagorinsky 1953; Hoskins and Karoly 1981). However, we believe it is possible that such a response may indeed be consistent, given a suitable basic state flow, and a suitable vertical distribution of thermal forcing, a vital component of which must be the transient eddy heat flux convergence (see below).

The purpose of the discussion below, then, is not to diminish the importance of transient eddy momentum forcing, but to suggest that the effects of thermal forcing cannot be discounted a priori.

The steady-state vorticity and thermodynamic equations on a β plane, linearized about a zonal mean basic state, $\bar{u}$, can be written

$$
\bar{u}_a \xi' / ax + \beta v' = -f_0 \nabla . v' \\
\bar{u}_a \theta' / ax + v' \bar{\theta} / ay + w' \bar{\theta} / az = Q'
$$

where $\xi'$ and $\theta'$ are the deviations of relative vorticity and potential temperature respectively about the zonal mean. Here $Q$ represents both the mean diabatic heating rate, and the total transient eddy heat flux convergence.

The ratio of the three terms on the left-hand side of the thermodynamic equation can be written as

$$
\bar{u}v' / H_Q : \bar{u}w' / H_u : f_0^{-1} w N^2
$$

using the thermal wind equation. Here $H_Q$ and $H_u$ are the height scales of the heat source and zonal velocity respectively.

If the mean wind is sufficiently weak at low levels, then the vorticity equation reduces to a Sverdrup balance: $\beta v' = -f_0 \nabla . v'$, from which one can obtain an estimate of vertical velocity: $w' \sim f_0^{-1} \beta H_Q \xi'$. Using this, the ratio of terms in the thermodynamic equation becomes $\gamma_Q : \gamma_u : 1$ where

$$
\gamma_Q = f_0 \bar{u} / (\beta N^2 H_Q H_u) \quad \text{and} \quad \gamma_u = f_0 \bar{u} / (\beta N^2 H_Q H_u).
$$
Hoskins and Karoly discuss the balance of terms when $\gamma_n \gg \gamma_0$ and $\gamma_n \gg 1$ below the heating maximum. In this case low-level heating is balanced by the advection of cool air from the north giving a downstream trough. The associated advection of planetary vorticity at low levels is balanced by vortex compression implying low-level divergence and descent over the heating region.

This balance does not apply to the GCM response, where we have noted strong low-level convergence. Furthermore for both atmosphere and GCM there appears to be an approximately equivalent barotropic positive height anomaly downstream of the s.s.t. anomaly.

Estimates of climatological diabatic heating (Lau 1979) show a mid-latitude maximum in the planetary boundary layer. Averaged over the Atlantic storm track a mean diabatic heating rate of about 1 or 2 K d$^{-1}$ is typical. White and Hoskins (to be published) have recently estimated typical transient eddy heat flux convergences along the Atlantic storm track. They find that the vertical heat flux convergence cannot be neglected, and that, combined with the horizontal convergence, the total effect of the (band-pass) transient eddies is to provide a cooling of about 1 K d$^{-1}$ on the mean flow in the lower troposphere, significantly offsetting diabatic heating; and a warming in excess of 1 K d$^{-1}$ in the upper troposphere with a maximum near 350 mb.

Hence it is difficult to assess the true height scale $H_Q$; however, with the influence of transient eddy heat flux convergence in mind, an estimate $H_Q \sim 4$ km may not be unreasonable. With a value $H_u \sim 4$ km near the RM area there is the possibility of some cancellation between the two horizontal advection terms in the thermodynamic equation. As discussed in section 2, there appears to be considerable cancellation between the advection of anomalous temperature by the mean wind, and advection of mean temperature by the anomalous wind, in the model's mid troposphere.

In terms of the linearized thermodynamic equation, then, the sum of the first two terms may be much less than $\gamma_0$ or $\gamma_n$, in which case vertical advection could not be ignored as a significant term to offset the effects of $Q$. This would require ascent over the thermal source, and low-level convergence. From the Sverdrup relationship the low-level vortex stretching induced by this convergence must be balanced by advection of planetary vorticity from the south, giving a downstream ridge.

With such a thermal forcing the dominant horizontal scale excited by the forcing will be close to the stationary 'free' Rossby wavelength, $2\pi \sqrt{u_0/\beta}$ where $u_0$ is some depth-averaged zonal mean wind. Above the level where $u = u_0$, zonal advection of relative vorticity will be larger than the meridional advection of planetary vorticity; below this level meridional advection will be larger. Such a balance can be achieved with an equivalent barotropic streamfunction, in contrast to the baroclinic response when the Sverdrup balance holds at all levels.

These ideas can be made more explicit by considering a two-layer $\beta$ plane model with zonal winds and densities $u_{1,2}$ and $\rho_{1,2}$, where '1' denotes the upper level and '2' the lower layer. We take the heating maximum to occur at the interface between the two levels. The stationary Rossby wavenumber, $K_s$, associated with the depth-averaged zonal mean wind is given by $K_s^2 = 2\beta/(u_1 + u_2)$. Now impose a heating anomaly

$$Q' = Q_0 \exp[i(kx + ly)] \quad |x| < 2\pi/k, \quad |y| < 2\pi/l,$$

where $K_s^2 = k^2 + l^2$, onto this basic state. In the thermal source region, the divergence

$$D' = D_{1,2} \exp[i(kx + ly)]$$

and geostrophic streamfunction

$$\psi' = i \psi_{1,2} \exp[i(kx + ly)]$$
satisfy the vorticity equation
\[ u_{1,2}k(k^2 + l^2)\psi_{1,2} - k\beta\psi_{1,2} = -f_0D_{1,2}. \]  
(1)

Now writing \( u_1 = \frac{1}{2}(u_1 + u_2) + \frac{1}{2}(u_1 - u_2) \), and similarly for \( u_2 \), (1) becomes
\[ \beta k \delta_u \psi_1 = -f_0D_1 \]  
(2a)
\[ \beta k \delta_u \psi_2 = f_0D_2 \]  
(2b)

where \( \delta_u = (u_1 - u_2)/(u_1 + u_2) \). From the continuity equation (with \( w = 0 \) at top and bottom level)
\[ \rho_1D_1 + \rho_2D_2 = 0. \]  
(3)

Dividing (2a) and (2b) and using (3)
\[ \psi_1/\psi_2 = \rho_2/\rho_1, \]  
(4)

an equivalent barotropic response. With low-level convergence \( (D_2 < 0) \), (2) implies that the meridional wind \( u_{1,2} > 0 \) giving a downstream ridge at both levels.

Let us now turn to the thermodynamic equation. The horizontal advection terms in the thermodynamic equation can be written as
\[ \bar{u} \frac{\partial}{\partial x} (\partial \psi'/\partial z) - (\partial \psi'/\partial x)(\partial \bar{u}/\partial z) = \bar{u}^2 \frac{\partial}{\partial x} \frac{\partial}{\partial z} (\psi'/\bar{u}). \]

In terms of the two-layer model the expression in brackets on the right of this equation will vanish provided that \( \psi_1/\psi_2 = u_1/u_2 \). From (4), this will occur if \( u_1\rho_1 = u_2\rho_2 \).

As mentioned above it is difficult to estimate the precise location in the vertical of the heating maximum, however, let us suppose this maximum occurs at \( z = 4 \text{ km} \). From Lau et al. (1981) we take \( u_1 = u(6 \text{ km}) \) near the RM area \( \sim 20 \text{ m s}^{-1} \); \( u_2 = u(2 \text{ km}) \) near the RM area \( \sim 10 \text{ m s}^{-1} \). Hence with \( \rho_1/\rho_2 = \rho(470 \text{ mb})/\rho(800 \text{ mb}) \sim 470/800 \)
\[ u_2\rho_2 : u_1\rho_1 \sim 10 \times 800 : 20 \times 470 = 0.85 \sim 1 \]

and adiabatic cooling by ascent will be important in balancing \( Q \). If this ratio is exactly equal to 1 then \( N^2w' = Q' \). From the continuity equation
\[ \rho w'|\delta = \int_0^{\Delta z} \rho D \, dz, \]
hence writing \( w' = w_0 \exp\{i(kx + ly)\} \) we have
\[ D_1 = Q_0(\rho_1 + \rho_2)/(2\rho_1 \Delta z N^2) \quad \text{and} \quad D_2 = -Q_0(\rho_1 + \rho_2)/(2\rho_2 \Delta z N^2) \]

where \( \Delta z \) is the layer thickness, giving upper-level divergence and lower-level convergence. From (2), the final solution is \( \psi_{1,2} = A \rho_1^{1/2} \)
\[ A = -(f_0/\beta k)(\delta_u/N^2 \Delta z)(\rho_1 + \rho_2)Q_0. \]

Two other aspects of the GCM and atmospheric responses need to be mentioned. Firstly, the negative height anomaly over Europe may possibly be explained in terms of downstream Rossby wavetrain propagation, as discussed by Hoskins and Karoly (1981). Secondly, we have noted in the GCM's lowest levels, a weak negative height anomaly directly over the s.s.t. anomaly. This is probably associated with the boundary layer parametrization. The model's boundary layer extends no higher than the second sigma level (at \( \sigma = 0.8 \)), and within it potential temperature is constrained to be height independent. Sensible and latent heat fluxes at the surface are given by conventional
bulk aerodynamic formulae. Hence, the imposition of a warm s.s.t. anomaly will lead to a warming throughout the boundary layer above it both by sensible and latent heating. By the hydrostatic approximation, a negative geopotential height anomaly, decreasing in magnitude from the surface to the top of the boundary layer, would be consistent with such a warming.

(b) The influence of the atmosphere on s.s.t. in the RM area

The response of North Atlantic s.s.t. to atmospheric forcing processes has been studied by Daly (1978). A heat energy equation, depth averaged over the ocean mixed layer, was used with given meteorological forcing from a number of case studies to provide values for the month-to-month change of mean monthly s.s.t. anomalies. One of Daly's case studies corresponds approximately to the conditions described in sections 2 and 3, with an anomalous anticyclonic surface pressure distribution over the mid North Atlantic for November and December 1966. The position of the centre of the anomaly corresponds fairly well with the model response in Fig. 5(a), and approximately with the centre of the observed composite anomalies in Figs. 11 and 12.

The thermodynamic equation in Daly's mixed layer model can be written as

$$\frac{\partial T'}{\partial t} = -\frac{\tau' \times \mathbf{k}}{\rho \chi h} \cdot \nabla_h T' - \left( \frac{\nabla_{\text{tot}} T'}{\rho \chi h} + \frac{\tau' \times \mathbf{k}}{\rho \chi h} \right) \cdot \nabla_h T' + \nabla_h \nabla^2 T' + \frac{H' + E'}{\rho \chi h}.$$  

The overbars refer to climatological values, and primes represent departures therefrom. Daly uses this equation in diagnostic form to calculate the change in monthly-mean anomalous temperature, $T'$, of the mixed layer, which is assumed to have constant depth, $h$. The terms on the right-hand side of the equation are calculated from observations for two consecutive months, and averaged to form a two-month mean. They are: (i) the advection of climatological mean ocean temperature $\bar{T}$ by the anomalous drift current, itself driven by the anomalous surface wind stress, $\tau'$; (ii) the advection of anomalous isotherms $T'$ by the total mean and anomalous ocean current; (iii) large-scale horizontal diffusion of anomalous s.s.t. by small-scale ocean eddies; and (iv) changes resulting from the effect of anomalous sensible, $H'$, and evaporative, $E'$, heat loss, determined by bulk aerodynamic formulae. Other terms have their conventional meanings.

Daly's model correctly diagnosed an anomalous warming $\partial T'/\partial t$ of 1 K/month between November and December at about 45°N 40°W, and the contributions due to the anomalous drift current and sensible heating are about the same. Relative to the anomalous pressure distribution, the s.s.t. anomaly develops to the SW of the high centre. The anomalous surface wind is approximately easterly over the s.s.t. anomaly, reducing the surface sensible heat loss, and driving an anomalous northward Ekman drift current which advects warm mean ocean temperatures from the south.

Since the positions of the surface anticyclones in Figs. 11 and 12 do not correspond exactly with Daly's case study, a quantitative comparison with his results cannot be made. Nevertheless one might expect that a south-easterly surface wind anomaly over the RM area would initially lead to a reduction in heat loss from the ocean surface, both by a reduction in westerly wind speed, and by an increase in the surface temperature of the air. Since the meridional temperature gradient is greater near the RM area than near Daly's surface anticyclone over the mid Atlantic (see Fig. 1(a)), it is possible that the advection of mean s.s.t. by the anomalous drift current may give rise to larger anomalies than in Daly's case.

It is possible, therefore, to envisage a steady-state situation whereby a substantial warm s.s.t. anomaly had developed and was losing sensible and latent heat into the
atmosphere at a rate which was balanced by the advection of warm water from the south by Ekman drift currents. Enhanced convection and latent heat release over the s.s.t. anomaly and anomalous baroclinic wave activity could then drive a surface wind distribution which would maintain the northward oceanic Ekman drift.

In the absence of a quantitative study using a coupled ocean–atmosphere model, it is only possible to speculate that air–sea interaction feedback may account for the observed persistence in autumn and early winter of s.s.t. anomalies in the RM area, and the atmospheric circulation perturbations described above. Nevertheless the arguments outlined above appear to suggest that this may not be unreasonable.

6. CONCLUSIONS

A modelling and observational study of the relationship between sea surface temperature in the north-west Atlantic and the northern hemisphere general circulation has been undertaken. The observational study was performed on two independent thirty-year periods, and results agreed well over the Atlantic and Europe. The modelling study comprised four pairs of 50-day integrations, each pair consisting of runs with warm and cold s.s.t. anomalies near the coast of Newfoundland. In the model the magnitude of the anomaly was enhanced by a factor of about 1.5 over Ratcliffe and Murray’s (1970) warm December composite (see Fig. 17). Allowing for this enhancement, the model and observations were in general agreement, with a positive geopotential height anomaly over the mid Atlantic and a negative geopotential height anomaly over Europe. In both cases the vertical structure was approximately equivalent barotropic. Four further pairs of integrations were run with halved s.s.t. anomalies, and still appeared to show positive height differences over the mid Atlantic, parts of which were just significant at the 10% level. The magnitude of these differences appears to depend approximately linearly on the magnitude of the s.s.t. anomaly.

The s.s.t. anomalies occur near the interface of the Gulf Stream and Labrador Current, under a region of intense cyclogenesis. GCM diagnostics and a simple two-layer model suggested that the momentum and thermal forcing by anomalous baroclinic wave activity was important in maintaining the time-mean geopotential height anomaly.

Results from the mixed layer model of Daly (1978) suggested that, both through anomalous Ekman drift currents and sensible heat fluxes, the s.s.t. anomalies are initially forced by the surface wind perturbations. It is speculated that in a steady state the s.s.t. anomaly loses sensible and latent heat to the atmosphere, which is balanced by northward advection of warm water driven by surface wind anomalies. In a future experiment we hope to test this idea quantitatively by applying the observed surface pressure perturbations found in this study, as boundary conditions for a North Atlantic mixed layer model.

The GCM response to these mid-latitude s.s.t. anomalies are certainly weaker than the response to tropical anomalies of similar magnitude (Palmer and Mansfield 1984, 1985a, b). Nevertheless, it is possible that positive air/sea interaction feedback could help the anomalies to develop and persist, particularly in late autumn and early winter.

ACKNOWLEDGMENTS

We thank Prof. B. J. Hoskins and Dr G. J. Shutts for useful discussions and Prof. J. Namias for encouraging and helpful correspondence.
REFERENCES


Corby, G. A., Rowntree, P. R. and Gilchrist, A. 1977

Daly, A. W. 1978

Folland, C. K., Chan, L. Y. and Maryon, R. M. 1982
‘Associations between some mid-latitude sea surface temperature anomalies in November and N. Atlantic surface pressure patterns in early December’. Met. O. 13 Branch Memorandum No. 120. Meteorological Office, Bracknell

Folland, C. K., Parker, D. E. and Kates, F. E. 1984

Horel, J. D. and Wallace, J. M. 1981

Hoskins, B. J. and Karoly, D. J. 1981

Hoskins, B. J., James, I. and White, G. 1983
The shape, propagation and mean flow interaction of largescale weather systems. *ibid.*, 40, 1595–1612

Response of a general circulation model to sea surface temperature perturbations. *ibid.*, 31, 857–868

Response of the NCAR general circulation model to prescribed changes in ocean surface temperature. Part I: Mid-latitude changes. *ibid.*, 34, 1200–1213

Lau, N.-C. 1979
Observed structure of tropospheric stationary waves and the local balances of vorticity and heat. *ibid.*, 36, 996–1016

Lau, N.-C., White, G. H. and Jenne, R. L. 1981
‘Circulation statistics for the extratropical northern hemisphere based on NMC analyses’. NCAR Technical Note 171

Lorenz, E. N. 1979

Minhinick, J. H. and Folland, C. K. 1984

Namias, J. 1964

Namias, J. and Roads, J. O. 1985

Palmer, T. N. and Mansfield, D. A. 1984
The response of two atmospheric general circulation models to sea surface temperature anomalies in the tropical east and west Pacific. *Nature*, 310, 483–485

Parker, D. E. 1980
Climatic change or analysts' artifice?—a study of grid-point upper-air data. *Met. Mag.*, 109, 129–152


Ratcliffe, R. A. S. and Murray, R. 1970

Smagorinsky J. 1953
The dynamical influence of large-scale heat sources and sinks on the quasi-stationary mean motions of the atmosphere. *ibid.*, 79, 343–366

Teleconnections in the geopotential height field in the northern hemisphere winter. *Mon. Wea. Rev.*, 109, 784–812