Oceanic Changes Associated with Global Increases in Atmospheric Carbon Dioxide: A Preliminary Report for the Atlantic Coast of Canada

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OCEANIC CHANGES ASSOCIATED WITH GLOBAL INCREASES IN ATMOSPHERIC CARBON DIOXIDE: A PRELIMINARY REPORT FOR THE ATLANTIC COAST OF CANADA

by

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Measurements over the past few decades clearly indicate that the concentration of atmospheric carbon dioxide and other so-called "greenhouse" gases is on the rise. Model studies indicate that this trend will result in a significant alteration of the Earth's climate during the next century. Expected changes include: a global increase in surface temperature especially in winter and at high latitudes; an intensified hydrological cycle including increased run-off at high latitudes; a decrease in the strength of winds in the lower atmosphere; a decrease in the areal coverage and thickness of ice in polar regions; an earlier seasonal snow-melt and sea-ice break-up; and a later freeze-up. In a scenario with the atmospheric concentration of CO_2 twice pre-industrial levels, predicted surface warming is of the order of a few degrees Celcius and associated changes in regional hydrological balances and wind forcing are of the order of 5 to 10 per cent of present day values. These changes could have far-reaching effects. This report was prepared at the request of the Atlantic Research Directors Committee of the Department of Fisheries and Oceans as a resource document to begin to address the question of how such changes might affect conditions in the waters off eastern Canada and hence the Atlantic Canadian fisheries industry.

Our basic approach has been to present our understanding of the dynamics which maintain the ocean in its present state and, based on this dynamical framework, to outline changes expected to accompany CO_2-induced climate changes predicted by atmospheric models. This approach does not fully account for the coupling between the atmosphere and the ocean, but it allows us to make reasonable predictions based on well-defined assumptions.

Some predictions for large-scale oceanic changes are relatively robust. Sea-surface temperatures will generally rise with the largest changes expected near 60°N; meridional gradients in surface salinity will increase creating fresher northern conditions and saltier subtropical conditions; and the thickness, areal extent and duration of sea-ice cover will decrease. Warmer and fresher northern surface waters and reduced wind mixing should generally result in thinner and more gravitationally stable high-latitude mixed layers. However, in regions where and during times when ice-cover is removed, enhanced mixing may locally counteract and even reverse the tendency towards thinner mixed layers. Reduced wind stress magnitudes should lead to weakened wind-driven gyres and an intensification of the hydrological cycle should lead to stronger buoyancy driven currents. Thus the Gulf Stream may relax somewhat while the Labrador Current may be enhanced. Reduced wind stress could also lead to a reduction in the cyclonic wind-driven circulation within the Labrador Sea and combined with more gravitationally stable mixed layers this could result in a reduction in the rate of deep water formation by convective processes in this region. As a response to warmer atmospheric temperatures, there may be a transient increase in iceberg production rates. The resulting addition of water to the ocean together with thermal expansion could raise sea levels by several tens of centimeters per century. These predicted oceanic responses would follow climatic change on decadal or shorter time scales and hence should be roughly synchronous with atmospheric changes. In addition there may be much slower adjustments of abyssal ocean properties to climatic change. An increase in the average value of evaporation minus precipitation over the Atlantic Ocean should result in an increase in basin-averaged surface salinity, and coupled with increased high-latitude air and sea-surface temperatures should result in saltier and warmer abyssal waters. Such adjustments would probably occur over several centuries.

The general high-latitude tendencies towards warmer and fresher ocean waters, and thinner and more seasonal duration of ice cover should also hold over the continental shelves off eastern Canada. However, shelf-water conditions should also be strongly influenced by changes in

the regional circulation. The expected increases in predipitation minus evaporation, and continental run-off in high latitudes should increase the strength and further reduce the salinity of the relatively cold and fresh, buyoyancy driven currents which account for the general southward flow over these shelves. These currents include the Labrador Current, which has a major influence on water properties over the entire eastern Canadian coastal region, and the outflow from the Gulf of St. Lawrence, which influences the Scotian Shelf and Gulf of Maine waters. The increased strength of these flows leads to the possibility that the shelf waters off eastern Canada may be warmed by a lesser amount than offshore waters at similar latitudes, but enhanced warming of the high-latitude source waters may offset this tendency. The tendencies towards reduced warming and increased freshening of shelf waters may be enhanced by the reduction in the strength and variability, and hence the influence, of the relatively warm and salty Gulf Stream. The increase in freshwater run-off should also result in small offshore displacements of the coastal and shelf-break fronts throughout the region, and the increased net solar radiation should lead to a reduction in the spatial extent of year-round tidally-mixed waters in areas such as the Gulf of Maine.

Associated with the warmer air and sea temperatures, there should generally be earlier seasonal snow-melts, spring thaws and ice break-ups; an increased occurrence of ice-free years in areas such as the Gulf of St. Lawrence; and a reduction in the seasonal extent of ice in areas such as the Labrador Shelf. As noted earlier, such changes in seasonal ice cover may allow increased wintertime mixing in these areas, opposing the general tendency towards thinner mixed layers.

We emphasize that these predictions are preliminary in nature. More definite statements will be possible with further development and understanding of fully coupled atmosphere-ocean-cryosphere models, detailed regional simulations, and validation of model predictions against relevant observations.

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Les mesures prises au cours des dernières décennies indiquent clairement que la concentration de gaz carbonique atmosphérique et d'autres gaz appelés "de serre" augmente. Des études de modèles indiquent que cette tendance aura pour résultat une importante altération du climat de la terre au cours du prochain siècle. Les changements attendus comprennent: une augmentation globale de la température en surface, spécialement en hiver et aux hautes latitudes; un cycle hydrologique intensifié avec augmentation du ruissellement aux hautes latitudes; une diminution de la force des vents dans la basse atmosphère; une diminution de la surface englacée et de l'épaisseur de la glace dans les régions polaires; une fonte des neiges saissonière et une débâcle de la glace de mer précoces, et un gel tardif. Dans ce scénario où la concentration atmosphérique de CO_2 est le double de celle de l'ère pré-industrielle, le réchauffement prévu en surface est de l'ordre de quelques degrés Celsius et les changements qui lui sont associés dans le bilan hydrologique régional et la force des vents sont de l'ordre de 5 à 10% des valeurs actuelles. Ces changements peuvent avoir des effects de portée incalculable. Ce rapport a été préparé à la demande du Comité des directeurs de recherche pour l'Atlantique du ministère des Pêches et des Océans comme document de base pour aborder la question concernant les effets éventuels de tels changements sur les eaux au large de la côte est du Canada et par conséquent, sur l'industrie des pêches canadiennes dans l'Atlantique.

Notre approche a consisté essentiellement à présenter notre interprétation de la dynamique qui maintient l'océan dans son état actuel et, à partir de ce cadre dynamique, à définir les changements qui accompagnent vraisemblablement les changements climatiques causés par le CO_2 et prévus par des modèles atmosphériques. Cette approche n'explique pas complètement la relation qui existe entre l'atmosphère et l'océan, mais nous permet de faire des prévisions raisonnables basées sur des hypothèses bien définies.

Certaines prévisions de changements océaniques à grande échelle sont relativement solides. Les températures superficielles de la mer augmenteront généralement et l'on s'attend aux plus grands changements aux environ de 60°N; les gradients méridionaux de salinité superficielle augmenteront, donnant une eau moins saline au nord et plus saline en milieu subtropical; l'épaisseur, l'étendue et la durée de couverture de glace diminueront. Les eaux de surface plus changés et moins salées au nord et la diminution du brassage par le vent auront pour résultat général des couches de mélange à haute latitude plus minces et plus stables gravitationnellement. Toutefois, dans les régions et pendant les périodes sans couverture de glace, un brassage plus marqué peut neutraliser et même renverser localement la tendance aux couches de mélange plus minces. La force réduite du vent devrait donner lieu à des tourbillons moins violents et une intensification du cycle hydrologique devrait conduire à des courants de poussée hydrostatique plus forts. Ainsi, le Gulf Stream pourrait diminuer, tandis que le courant du Labrador pourrait s'intensifier. La force réduite du vent pourrait aussi conduire à une réduction de la circulation cyclonique due au vent dans la mer du Labrador et, combinée à des couches de mélange plus stables gravitationnellement, elle pourrait entraîner une réduction de la vitesse de formation de l'eau profonde par des processus convectifs dans cette région. A cause du réchauffement de l'atmosphère, it se pourrait que la production d'icebergs augmente temporairement. L'addition résultante d'eau dans l'océan ainsi que l'expansion thermique pourraient entraîner une montée du niveau des mers de plusieurs dizaines de centimètres par siècle. Ces réponses océaniques prévues surviendraient une dizaine d'années ou moins après les changements climatiques, donc plus ou moins au moment des changements atmosphériques. De plus, il est possible que les propriétés abyssales des océans s'ajustent beaucoup plus lentement aux changements climatiques. Une augmentation de la valeur moyenne de l'évaporation moins les

précipitations au-dessus de l'océan Atlantique aurait pour résultat une augmentation de la salinité superficielle moyenne du bassin entier et, combinée à une augmentation des températures de l'air aux hautes latitudes et de la surface de la mer, rendrait les eaux abyssales plus salées et plus chaudes. De tels ajustements s'étaleraient probablement sur plusieurs siècles.

La tendance générale, vers des eaux océaniques plus chaudes et moins salées aux hautes latitudes, vers des couches de mélange superficielles plus stables et plus minces, et vers une diminution de l'étendue et de la durée saisonnière de la couverture de glace devrait aussi s'appliquer au-dessus des plates-formes continentales de l'est du Canada. Toutefois, les conditions de l'eau de la plate-forme devraient aussi être très sensibles aux changements dans la circulation régionale. L'augmentation prévue de l'écart entre les précipitations et l'évaporation, et l'écoulement continental aux hautes latitudes devraient augmenter la force et réduire davantage la salinité des courants d'eau relativement douce et froide, qui sont causés par la poussée hydrostatique et qui expliquent la circulation générale vers le sud sur ces plates-formes. Ces courants comprennent le courant du Labrador qui a une importante influence sur les propriétés de l'eau de toute la région côtière de l'est du Canada et le courant du golfe du Saint-Laurent qui influe sur les eaux du plateau Scotian et du golfe du Maine. A cause de la force accrue de ces courants il est possible que l'eau des plates-formes sur la côte est du Canada puisse être moins réchauffée que les eaux du large à des latitudes semblables, mais que le réchauffement plus marqué des eaux provenant des hautes latitudes puissent annuler cette tendance. La tendance vers une réduction du réchauffement et de la salinité des eaux des plates-formes pourrait être renforcée par la réduction de la force et de la variabilité, et donc de l'influence, du Gulf Stream dont l'eau est relativement chaude et salée. Le débit accru d'eau douce de ruissellement pourrait aussi avoir pour résultat de petits déplacements vers le large du front de la côte et de la plate-forme dans toute la région, et l'augmentation du rayonnement solaire net devrait conduire à une réduction de l'étendue spatiale des zones permanentes de mélange dû aux marées telles que le golfe du Maine.

Associés aux températures plus chaudes de l'air et de la mer, la fonte saisonnière de la neige, le dégel printanier et la débâcle saisonnière devraient avoir lieu plus tôt, le nombre d'années sans glace devrait augmenter dans des zones telles que le golfe du Saint-Laurent, et l'étendue saisonnière de la glace dans des régions telles que la plate-forme du Labrador devrait diminuer. Comme il a été mentionné plus tôt, de tels changements dans la couverture de glace saisonnière pourraient favoriser le brassage en hiver dans ces zones, brassage qui s'opposerait à la tendance générale vers des couches de mélange plus minces.

Nous insistons sur le fait que ces prévisions sont préliminaires. Des déclarations plus définitives seront possibles lorsque nous aurons perfectionné et mieux compris des modèles atmosphère-océan-cryosphere entièrement intégrés, effectué des simulations régionales détaillées, et validé les prévisions des modèles par comparaison avec des observations pertinentes.

1. INTRODUCTION

It has been realized for decades that man might influence the Earth's climate through the production of carbon dioxide (Callendar, 1938). Today there is clear evidence that the CO_2 content of the Earth's atmosphere is increasing (Fig. 1.1) and the WMO/ICSU/UNEP group of experts (World Meteorological Organization, 1981) has estimated that, by the year 2025, the atmospheric CO_2 concentration will be 450 (±40) parts per million, about 50% higher than pre-industrial values. It appears then that the CO_2 concentration will double sometime in the next century. The general consensus among experts is that a doubling of atmospheric CO_2 will result in an equilibrium global surface warming of approximately 3 (±1.5)°C (CO_2/Climate Review Panel, 1982) and increased concentrations of radiatively active trace gases other than CO_2 may substantially amplify this effect (Ramanathan <u>et al.</u>, 1985). Although this view is not unanimous (Idso, 1982), phenomena such as the 1982-83 El Nino serve as vivid reminders that the consequences of climate perturbations can be too devastating to be simply ignored.

Our aim here is to examine possible changes in the marine environment which may influence the Atlantic Canadian fisheries industry. Our starting point is the estimate of changes in atmospheric conditions obtained from atmospheric general circulation models; we use these results in turn to speculate on possible oceanic changes, particularly in the North Atlantic Ocean. It is important to realize at the outset that any discussion of changes in the ocean caused by the predicted changes in atmospheric forcing is bound to be uncertain. The most advanced models presently used to predict such changes are only beginning to include ocean dynamics in a detailed way, although the ocean is a critical part of both the physical

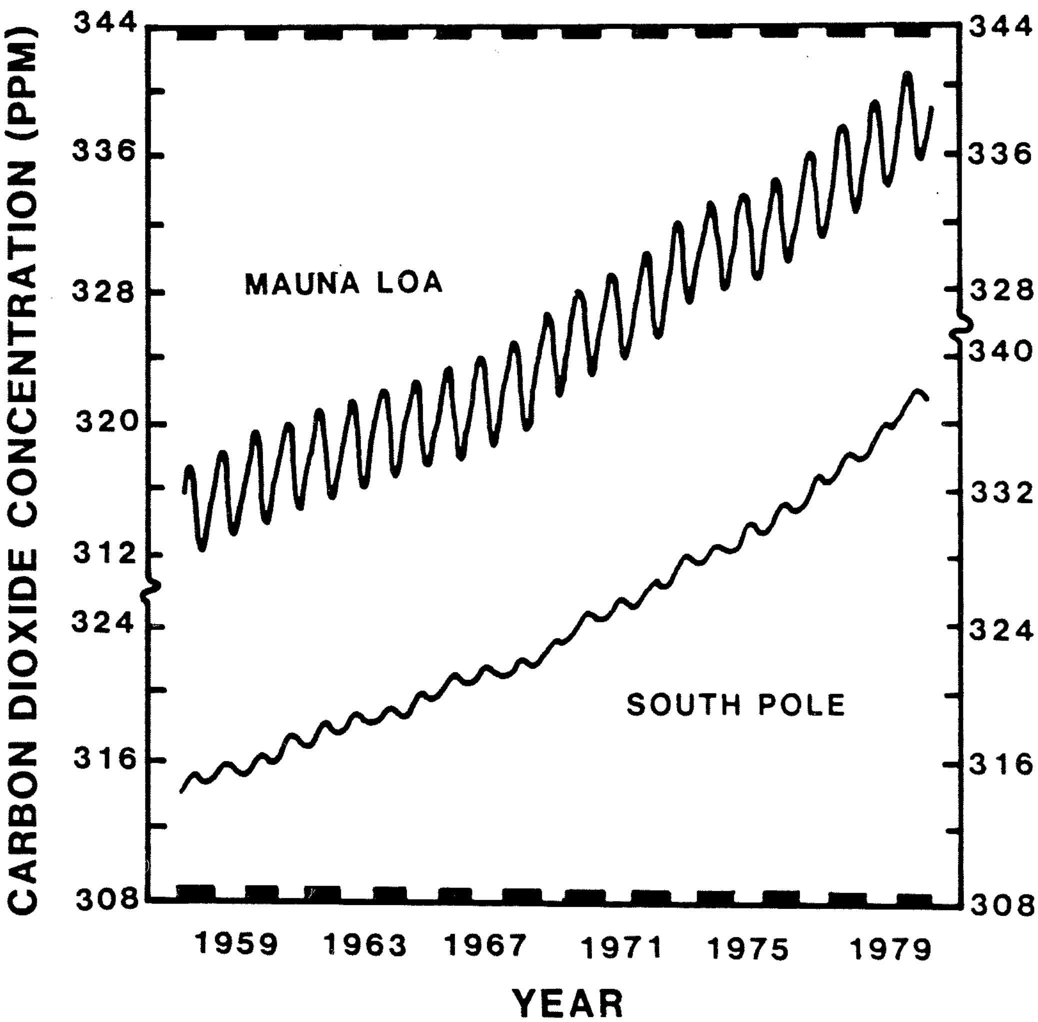


Fig. 1.1. Observations of atmospheric CO_2 concentration at Mauna Loa, Hawaii and at the South Pole. (Figure taken from Idso (1982)).

climate system and the geochemical processes controlling overall atmospheric CO_2 concentration. Thus, to isolate the oceans from the overall system and enquire about changes due to some predicted atmospheric change is clearly a questionable undertaking.

The general approach adopted in this report is to describe the ocean as it is today, present our understanding of the dynamics that determine this state, and then, based on this dynamical framework, put together a scenario of possible tendencies associated with predicted atmospheric changes. If either different atmospheric changes or different ocean dynamics are assumed, this scenario will change. Although we have alluded to some alternative possibilities, in general we have attempted to assimilate available information to give a "best guess" scenario together with a dynamical framework which should assist in formulating others.

In the next section we briefly summarize the changes in atmospheric conditions predicted for a doubling of atmospheric CO_2 concentrations. In Section 3 possible changes in the large-scale temperature and salinity structure of the North Atlantic Ocean are considered and in Section 4 possible changes in sea-ice conditions are discussed. Section 5 considers the wind-driven circulation of the North Atlantic Ocean. In Section 6 we speculate on how the physical oceanographic conditions on the continental shelves may change. Finally, a brief summary of our expectations is presented in Section 7, with a discussion of some of the problems facing modellers who attempt to predict changes in the Earth's atmosphere-ocean-cryosphere system.

2. ATMOSPHERIC RESPONSE TO CHANGES IN CO_2 CONCENTRATION

In the absence of an atmosphere, the Earth's equilibrium surface temperature would be determined by a balance between the absorption of incoming solar (short-wave) radiation and the radiation of infrared (long-wave) energy back into space:

$$\pi R^2(1 - \alpha)S_0 = 4\pi R^2 \sigma T_e^{\,4}$$

or

$$T_e = [S_0(1 - \alpha)/4\sigma]^{1/4} \qquad (1)$$

where T_e is the mean equilibrium surface temperature of the Earth,

R is the Earth's radius,

α is the albedo of the Earth,

S_0 is the flux (input/unit area) of incoming solar radiation,

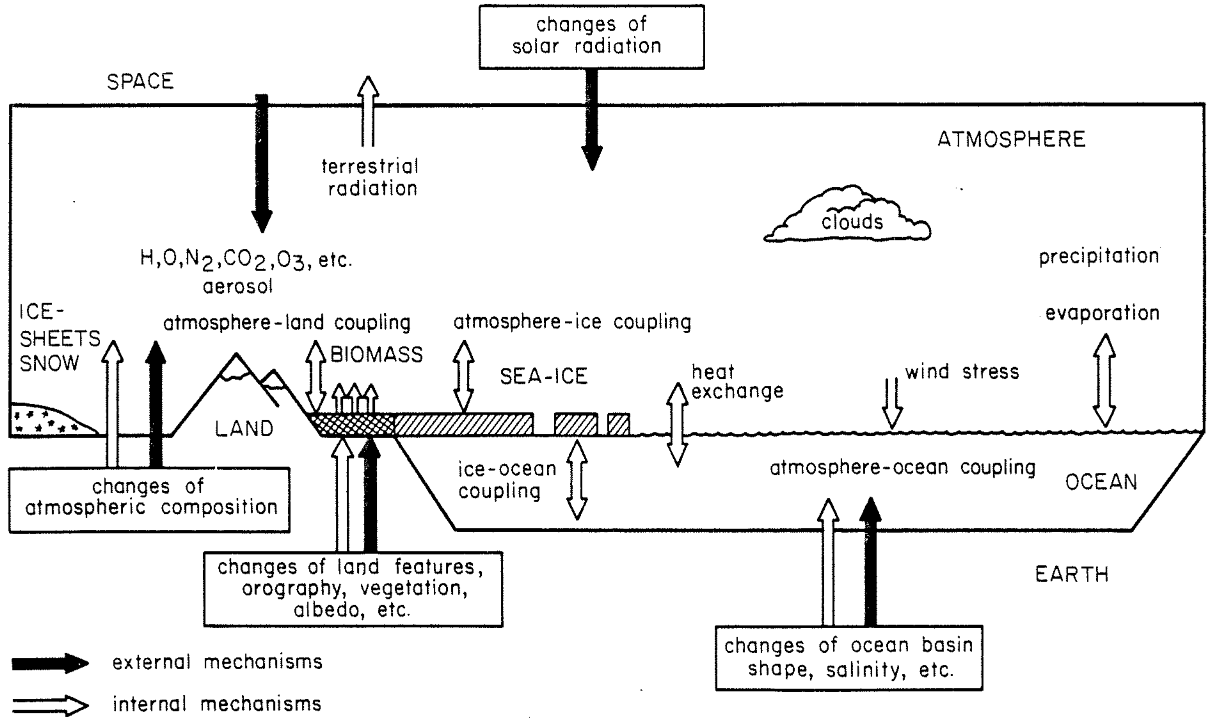
and σ (= 5.67×10^{-8} W m^{-2} K^{-4}) is a constant.

For a global mean albedo of 0.3 and a constant solar flux of 1367 W m^{-2}, (1) yields $T_e \approx 255$ K (-18°C).

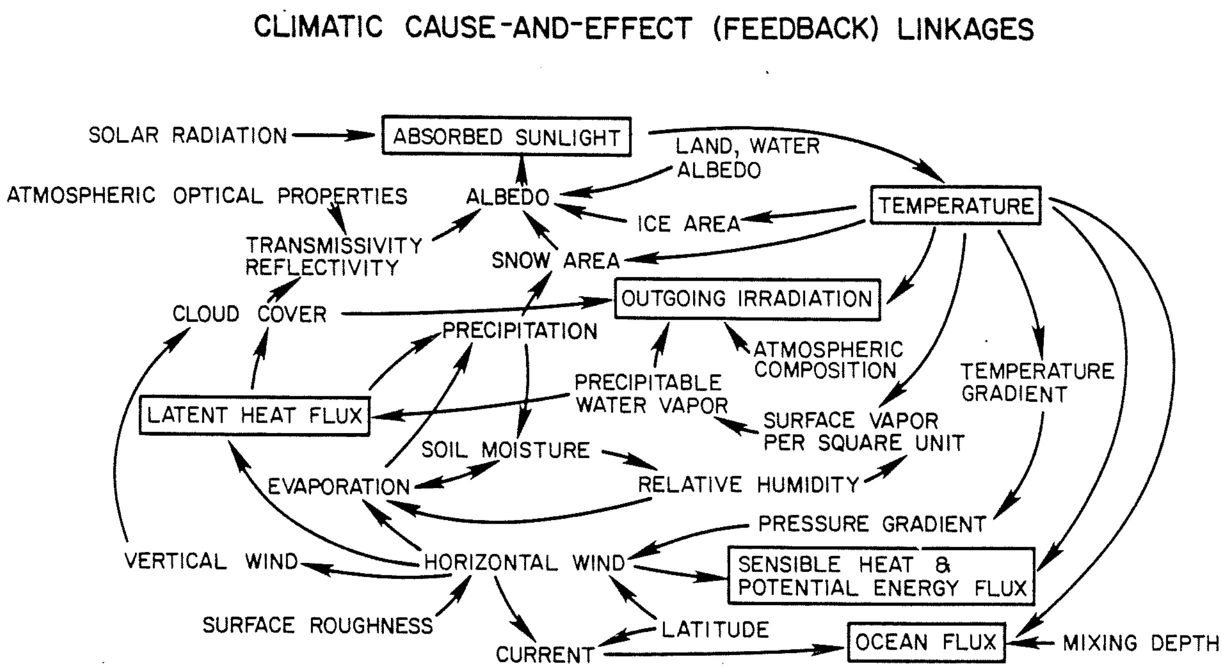
However, the Earth's atmosphere is fairly opaque to long-wave radiation and the effective height at which the outgoing radiation is in balance with incoming radiation is in fact about 6 km above the Earth's surface. Thus (1) implies that it is the temperature at this height which is approximately 255 K. Further, the opacity of the atmosphere to long-wave radiation causes the lower levels of the atmosphere to be heated preferentially; the excess heat is then transported upwards predominantly by convective instability. The decrease in pressure with altitude (and the influence of moisture) then implies that the temperature decreases away from the surface at a rate of about 5 to 6 K km^{-1} [less than the dry adiabatic lapse rate ($g/C_p \approx 10$ K km^{-1}) due to the release of latent heat by condensation], so that the Earth's equilibrium surface temperature is some 30 K higher than that expected from the radiative balance expressed by (1).

Since CO_2 strongly absorbs and re-emits long-wave radiation in the wavelength range of ≈12 to 18 μm, increasing the amount of CO_2 in the atmosphere will increase its opacity to long-wave radiation and hence increase the effective height at which the radiative balance (1) holds, which in turn will result in an increase in the surface temperature of the Earth. This is the basic process which has been termed the CO_2 greenhouse effect, though the analogy is clearly limited. As Hansen *et al.* (1981) point out, the surface temperature changes due to increasing the atmospheric CO_2 concentration are more analogous to the rise in surface level in a leaky bucket being filled with water when the size of the holes is slightly reduced (see Appendix for an extension of this analogy to include a crude ocean).

The above argument gives reasonable estimates of the Earth's mean surface temperature but it clearly neglects many complex and poorly understood interactions (Fig. 2.1). Within the context of the simple model discussed above, these secondary effects influence the equilibrium surface temperature of the Earth via changes in the albedo, changes in the effective height of radiative balance and changes in the lapse rate (i.e. the rate of decrease of temperature away from the Earth's surface). For example, increased moisture in the air will increase the effective height of radiative balance through increased absorption and re-emission of long-wave radiation (a positive feedback) and reduce the lapse rate through the release of latent heat as air parcels rise (a negative feedback); increased low-level clouds will increase the effective height of radiative balance



The components of the coupled atmosphere-ocean-ice-earth climatic system. The full arrows (→) are examples of external mechanisms, and the open arrows (⇒), examples of internal processes in climatic change. (After Living With Climatic Change, *Science Council of Canada)*



(Source: Schneider, 1976)

Fig. 2.1: Illustrations of the complexity of the problem with which climate modellers are faced. (Figure taken from F.P. Bretherton's (1978) article in <u>Oceanus, 21</u>).

through absorption and re-emission of long-wave radiation (a positive feedback) and increase the Earth's albedo through reflection of incoming short-wave radiation (a negative feedback); and increased high-level clouds will similarly increase the effective height of radiative balance (a positive feedback). Such feedback processes clearly complicate climate modelling. In spite of the resulting uncertainties about predictions obtained from present numerical models of atmospheric circulation, these models appear to be the most promising predictors of the effects of changes in atmospheric CO_2 concentrations.

Figure 2.2 shows examples from Manabe and Stouffer (1980) of the latitude-height distribution of the difference in zonal mean surface air temperature between two model runs, one with four times the present-day concentration of atmospheric CO_2 (4 X CO_2) and a control run with present-day levels (1 X CO_2). This particular model has realistic geography, seasonal variations and a mixed-layer ocean. The model ocean participates in the hydrological cycle and the seasonal heat balance, but does not include any heat transport by ocean currents. The warming of the lower levels of the atmosphere in the increased CO_2 scenario is apparent with the greatest warming at high latitudes and during winter (see Section 4 for a discussion of how ice influences temperature changes).

A second reasonably well-established prediction of these numerical general circulation models is an amplified hydrological cycle in the increased CO_2 scenaries. As might be expected under warmer conditions, evaporation increases at all latitudes resulting in balancing increases in precipitation (Fig. 2.3a). In the Northern Hemisphere the increased precipitation occurs primarily at latitudes above 45°N, resulting in a clear increase in precipitation minus evaporation (P-E) in this region and a

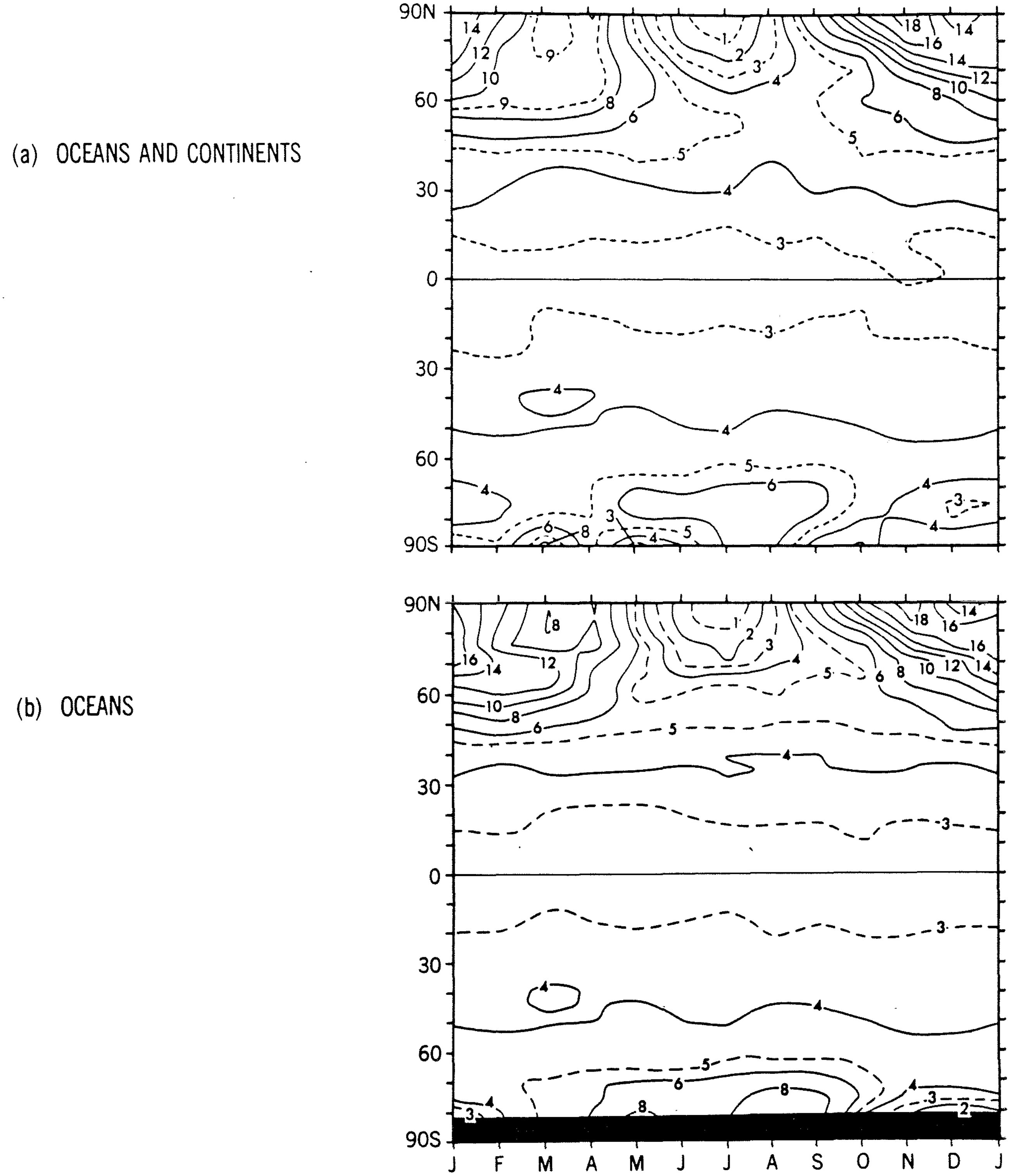


Fig. 2.2. Model predictions of the latitude-time distribution of zonal mean difference in surface air temperature (degrees Kelvin) between 4 x CO_2 and 1 x CO_2 experiments. Results are shown (a) for averages over both oceans and continents; and (b) for averages over ocean areas only. Corresponding differences for the 2 x CO_2 case would be roughly half these values. (Figure taken from Manabe and Stouffer, 1980.)

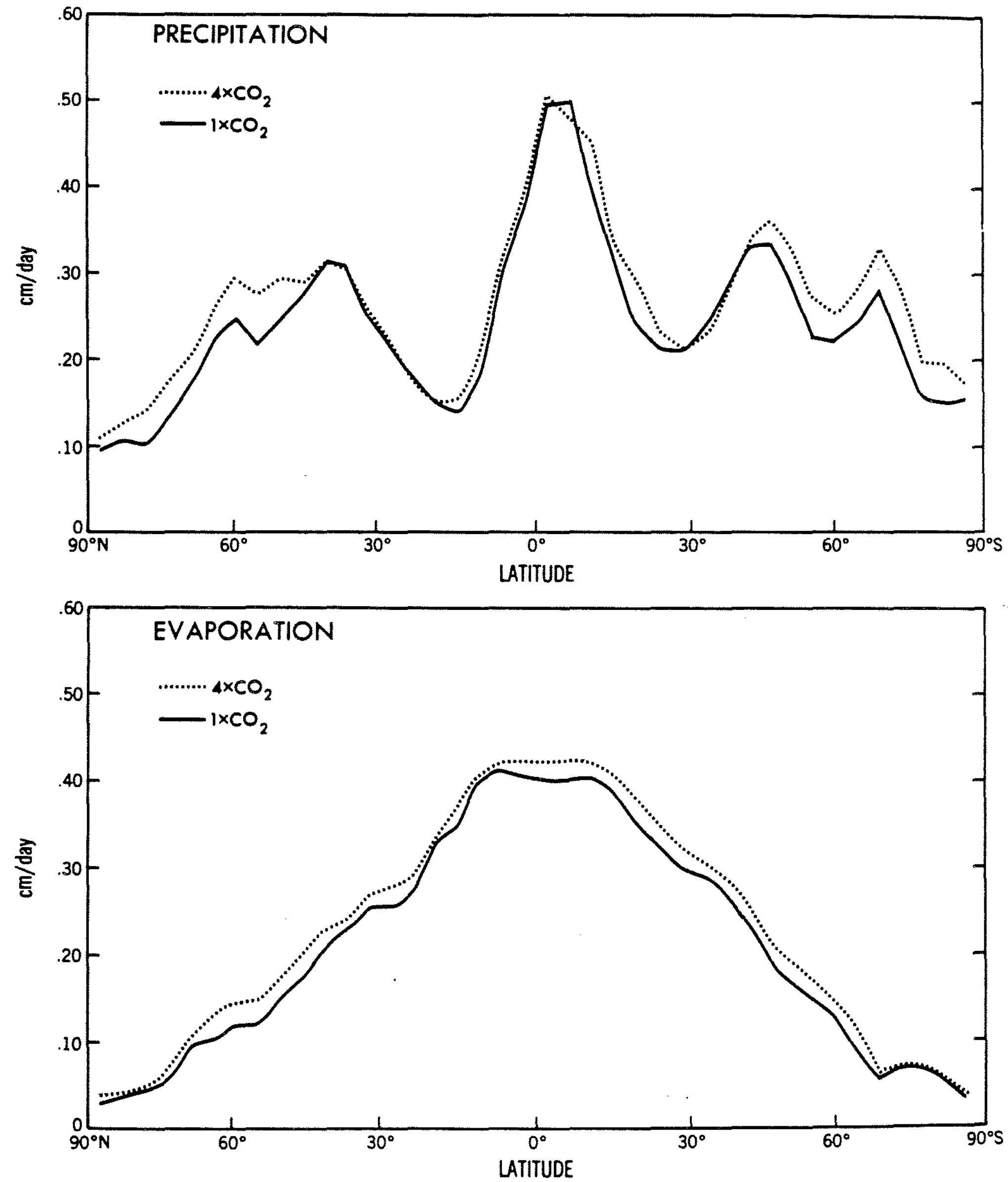


Fig. 2.3a. Model predictions of the latitudinal distributions of zonal mean precipitation and evaporation rates (cm/day). Solid and dashed lines are for 4 x CO_2 and 1 x CO_2 experiments respectively. Changes for 2 x CO_2 would be roughly half those for 4 x CO_2. (Figure taken from Manabe and Stouffer, 1980.)

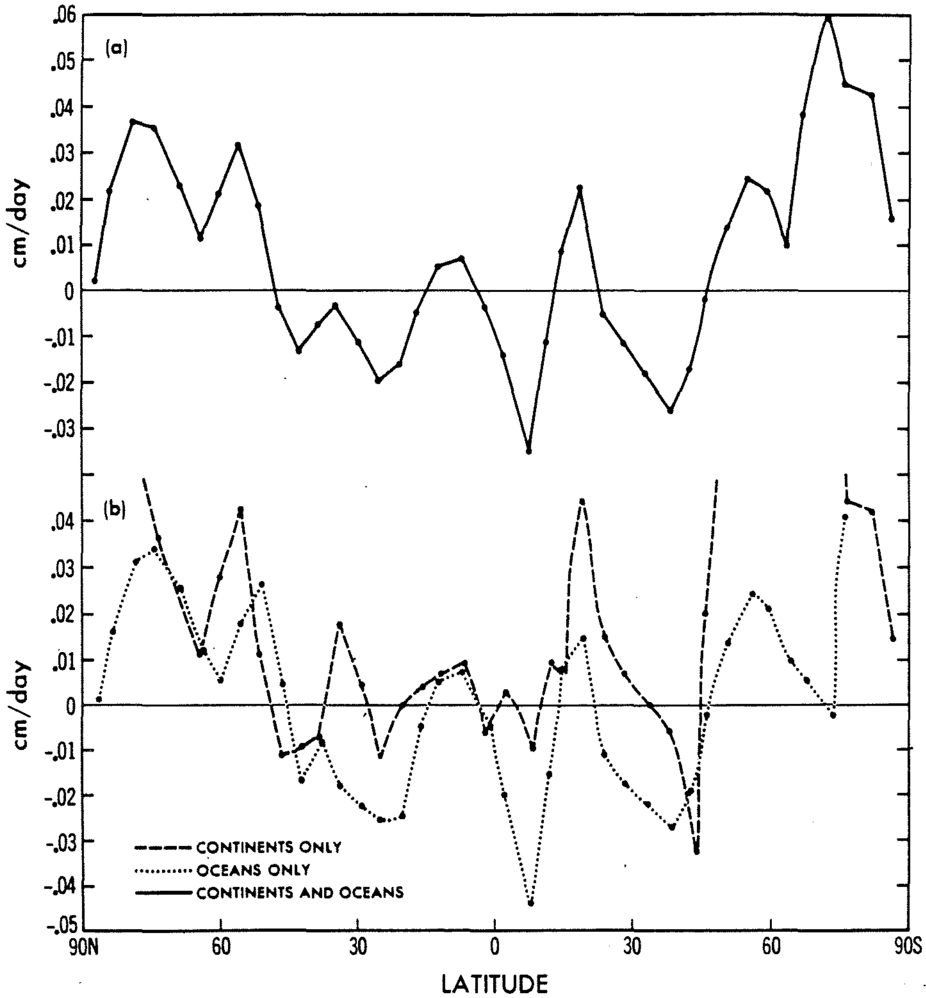


Fig. 2.3b. Model prediction of the latitudinal distribution of zonal mean difference in the annual mean value of precipitation minus evaporation between the 4 x CO_2 and 1 x CO_2 experiments (cm/day). Changes for 2 x CO_2 would be roughly half of those for 4 x CO_2. (Figure taken from Manabe and Stouffer, 1980.)

reduction in (P-E) at mid-latitudes (Fig. 2.3b). The increase in (P-E) at high northern latitudes results in a general increase in continental runoff north of 60°N, and there is a shift in time of the maximum runoff rate at these latitudes because of earlier snowmelt in a warmer climate. In the 4 x CO_2 scenario, the maximum runoff rate between 60° and 80°N is about one month earlier than for the 1 x CO_2 case. For a 2 x CO_2 scenario, changes are expected to be about half of the 4 x CO_2 results, leading to increases in precipitation, evaporation and (P-E) of order 5 percent, with maximum run-off at high latitudes occurring perhaps 2 weeks earlier on average.

Finally, we require some estimate of changes in surface wind stress acting on the ocean in a system with increased CO_2. Wind stress has not been emphasized in the presentation of model results to date, and predictions may not be reliable (CO_2/Climate Review Panel, 1982). However, there is some evidence that wind stress will be reduced by the influence of doubled CO_2 concentrations. The reduced meridional temperature gradient due to enhanced surface warming at high latitudes suggests that there may be an accompanying reduction in the large-scale winds. Manabe and Wetherald (1980) also present results which show a reduction in atmospheric eddy-kinetic energy at the surface by about 10% between 20 and 55°N. In the absence of further information we assume a doubled CO_2 scenario in which the magnitude of both the mean and eddy wind stress fields are reduced on the order of 10% over the entire North Atlantic without any associated change in pattern. This choice provides a basis for a discussion of ocean circulation change. Lacking any reliable results on changes in wind patterns, we proceed on the assumption that there are no associated changes in these patterns. It is certainly conceivable that wind stress patterns will also be altered by a doubling of atmospheric CO_2

and this could lead to substantially different results.

3. OCEANIC RESPONSE: A BROAD-BRUSH PICTURE

In this section we speculate on how the large scale thermohaline properties of the oceans (particularly the North Atlantic Ocean) will change in response to the predicted atmospheric changes reviewed in the previous section. For the purpose of this discussion the ocean may be divided into four basic regions: the surface mixed layer, the mid- to low-latitude region below the mixed layer but above the base of the permanent thermocline, the high-latitude region below the mixed layer which is renewed by convection, and the abyssal waters below the thermocline (Fig. 3.1).

The mixed layer is the region immediately below the ocean surface which is in direct contact with the atmosphere. It adjusts to atmospheric changes on a time scale of order 5 years (Bryan et al., 1984) and hence will react to climate change relatively rapidly. Assuming a mixed layer which is 70 m thick (a typical mixed layer depth used in atmospheric General Circulation Models), the heat capacity of this layer is more than twenty times that of the entire overlying atmosphere (at 10°C) and it represents an essentially infinite source of moisture. More than 80 per cent of global evaporation and precipitation actually occur over the oceans (Dietrich et al., 1980), and this hydrological cycle is a vital part of the atmosphere's general circulation. For these reasons, at least a mixed-layer ocean model is generally included in atmospheric general circulation models. Although oceanic heat transport is not included in mixed-layer models, the resulting predictions of changes in mixed layer temperature are still useful estimates of the order of magnitude of changes expected. Figure 3.2 illustrates the changes in surface water temperature predicted

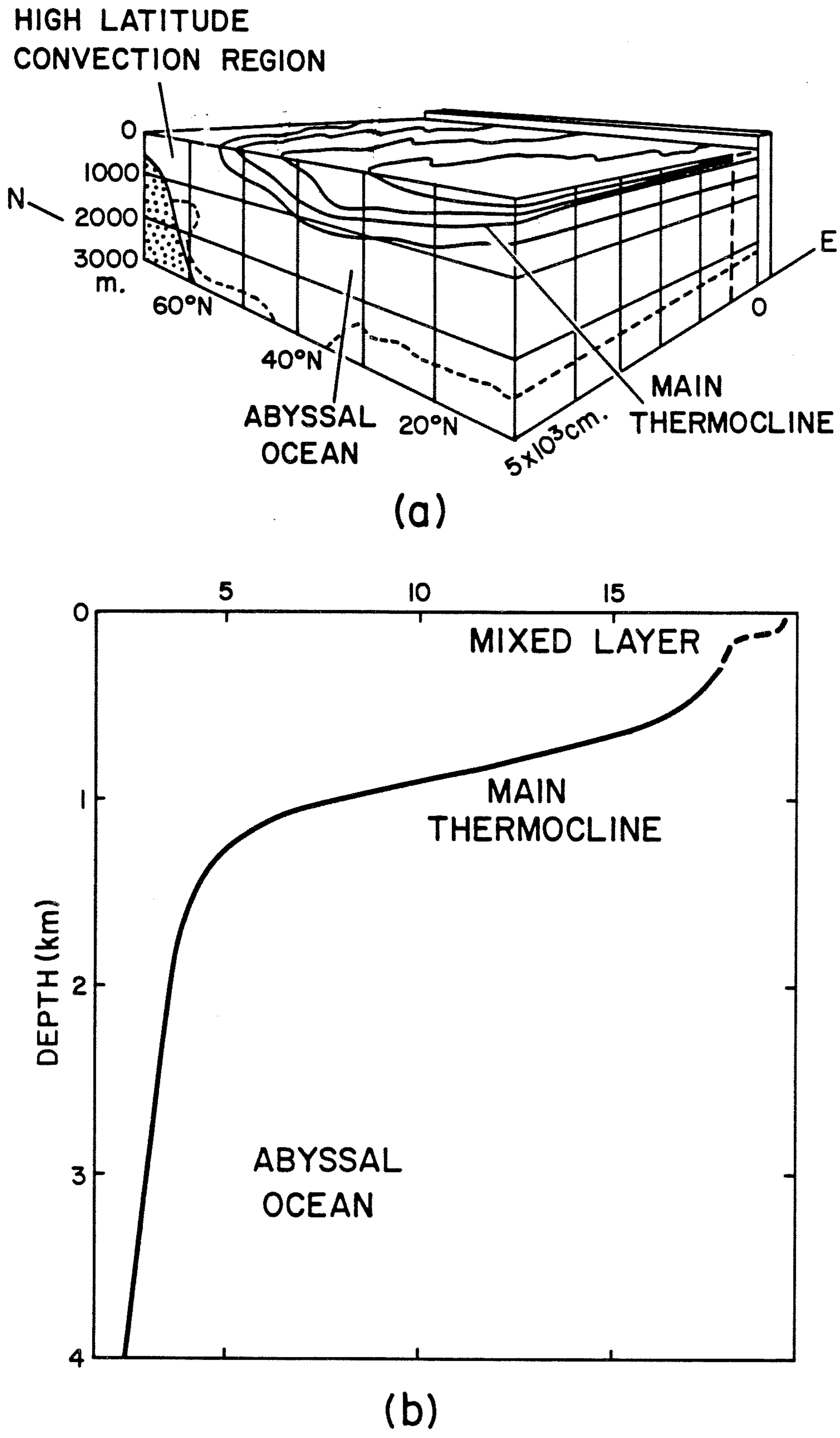


Fig. 3.1. (a) A schematic view of the temperature distribution in the North Atlantic. (b) Characteristic distribution of temperature with depth. (Figure adapted from Pedlosky, 1979.)

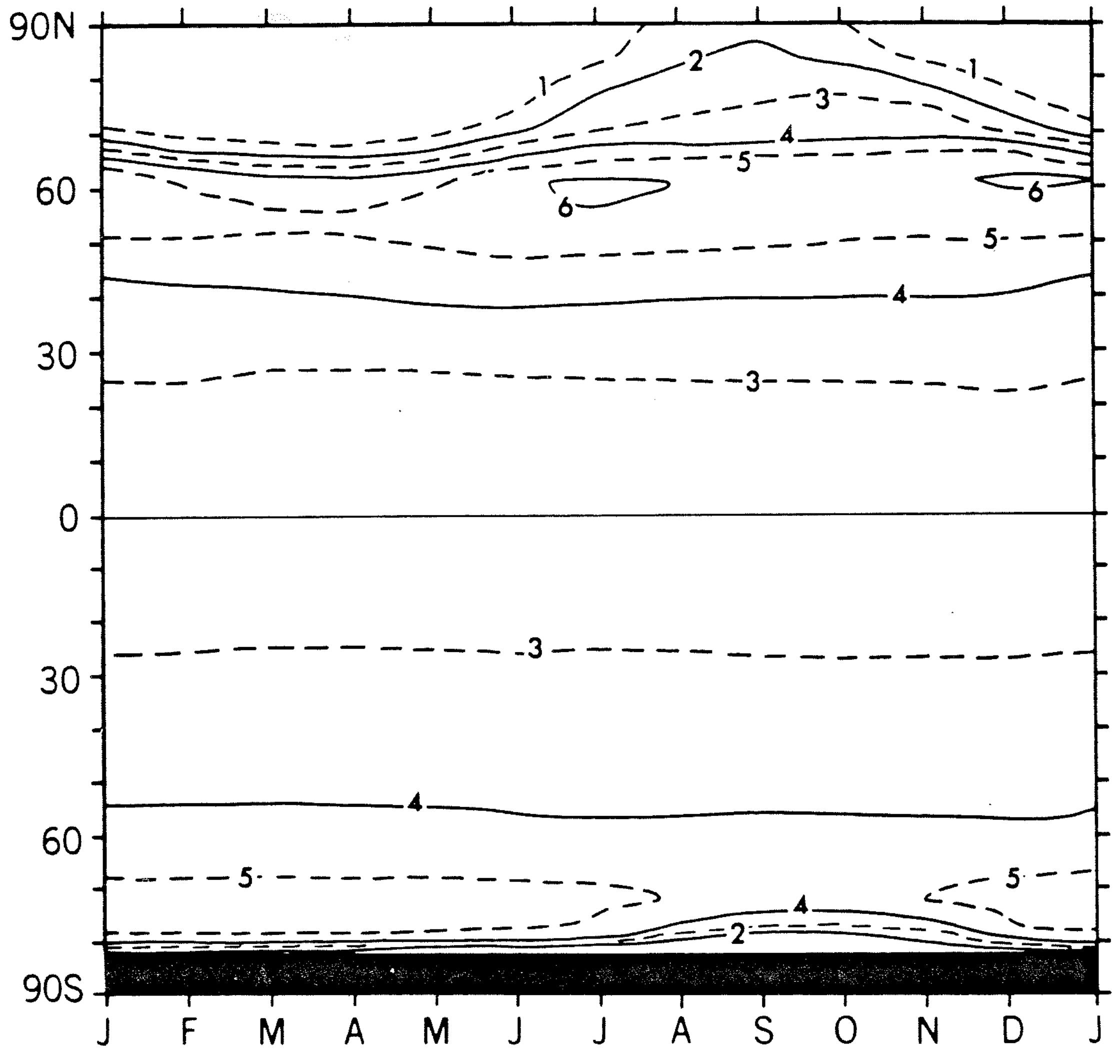


Fig. 3.2. Model prediction of the latitude-time distribution of the difference in zonal mean surface water temperature (°C) of the mixed layer between the 4 x CO_2 and 1 x CO_2 experiments. The difference will be roughly halved for the 2 x CO_2 case. (Figure taken from Manabe and Stouffer, 1980.)

by the Manabe and Stouffer (1980) model for 4 x CO_2. As expected the surface temperatures increase with changes of up to 6°C. The corresponding increases in temperature for doubled atmospheric CO_2 are expected to be of the order of 1-3°C. The pattern of oceanic temperature rise mirrors the poleward increase in atmospheric temperature change except in high latitudes where sea ice cover insulates the ocean from the overlying atmosphere (see Section 4). The largest increases in mixed layer temperature are in fact associated with a general poleward retreat of sea ice cover in the Arctic and Sub-Arctic. Seasonal temperature variations in the northern sector are also related to sea ice, with a pronounced summertime maximum warming in areas which had year round ice cover in the 1 x CO_2 scenario but become ice free in summer in the increased CO_2 model run. A second maximum warming in early winter reflects the delay in freeze-up associated with the generally warmer climate in the increased CO_2 scenario.

Since salinity does not directly affect atmospheric conditions, it is not normally predicted by atmospheric climate models. However, relatively small changes in salinity may affect not only the static stability of the water column but also the marine biosphere and hence it is important to examine possible changes. In this section we shall be concerned primarily with the salinity of the open-ocean North Atlantic. Changes in salinity over the continental shelves are discussed in Section 6.

The present-day distribution of near-surface salinity in the North Atlantic Ocean exhibits strong regional variations (Fig. 3.3). The large scale picture shows high salinities (≈ 37 ppt) at low- to mid-latitudes, decreasing somewhat towards the equator (to about 36 ppt) and decreasing substantially towards high latitudes (to less than 35 ppt in the Labrador Sea). This variation in surface salinity bears a strong resemblance to the

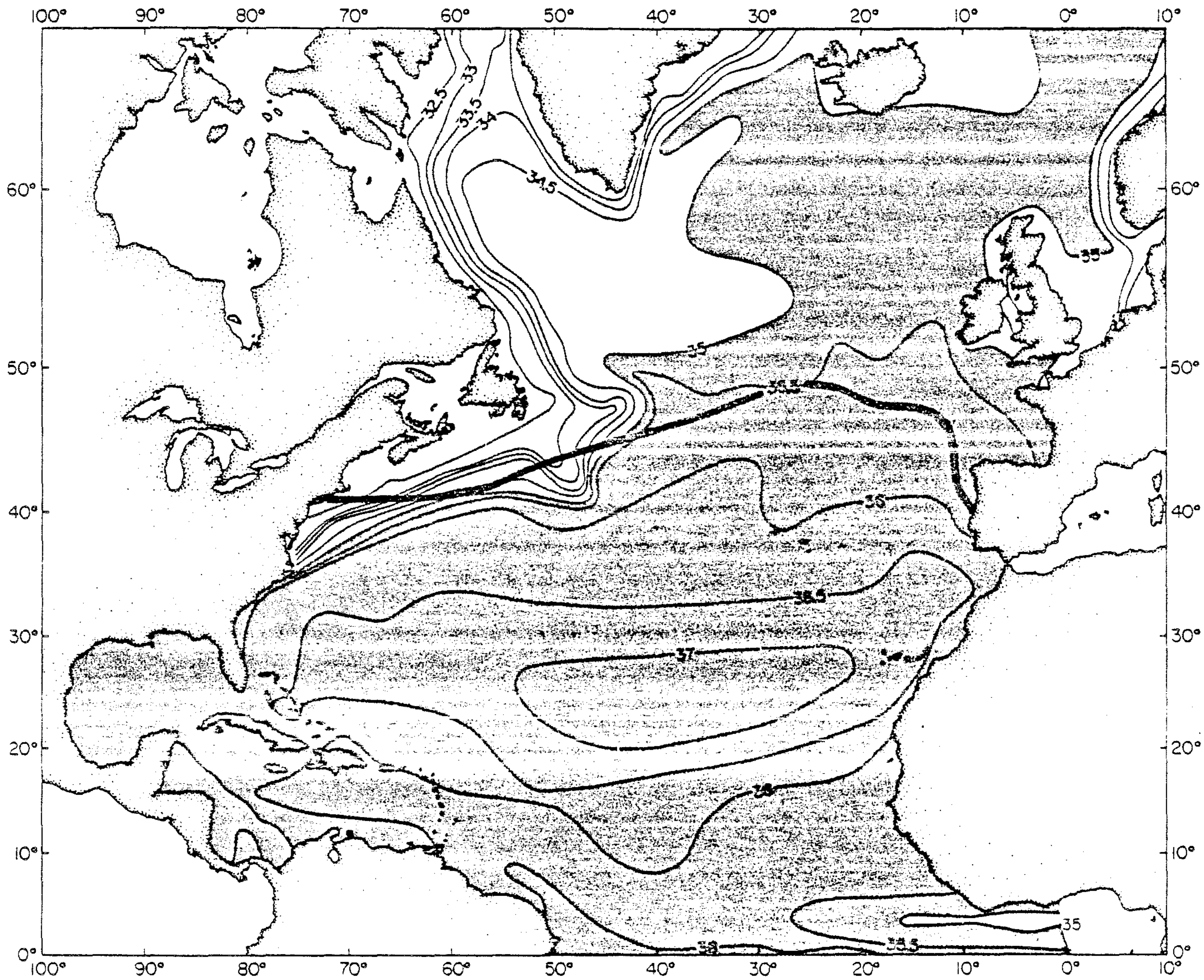


Fig. 3.3. The salinity in parts per thousand at a depth of 30 m in the North Atlantic Ocean. The heavy solid line running across the Atlantic is the zero in (P-E) taken from Plate 21 of Baumgartner and Reichel (1975). (Figure adapted from Worthington, 1976).

north-south variation in (E-P) over the region. Figure 3.4 from Dietrich et al. (1980) shows the correlation between north-south variations in global zonal-mean surface salinity and north-south variations in zonally averaged (E-P). In the North Atlantic, the surface salinity is also strongly influenced by an Arctic input of relatively fresh water, which is then carried southwards down the east coast of North America by the regional ocean circulation (see Section 6).

The more energetic hydrological cycle associated with a warmer lower atmosphere in a doubled CO_2 scenario should give increased (E-P) in sub-tropical latitudes which presently give up fresh water to the atmosphere on average. Balancing increases in (P-E) are expected for equatorial and polar regions already receiving a net input of fresh water from the atmosphere. The resulting tendency should be towards higher surface salinity in subtropical regions and lower surface salinity in polar and equatorial regions, with the final equilibrium depending on the ocean currents which will necessarily change to balance the new surface fluxes of fresh water. In crude terms, a 5 per cent increase in the amplitude of the hydrological cycle might be expected to change the surface salinity in the North Atlantic by amounts of order ± 0.05 ppt, given that the actual variations in present day surface salinity are of order ±1 ppt (Fig. 3.3).

On the western side of the North Atlantic Ocean, changes in the Arctic contribution of fresh water and local increases in continental run-off should maintain and even amplify the zonal trend towards fresher surface waters. A relatively fresh Arctic outflow results from net (P-E), continental run-off and inflow of relatively fresh surface waters from the northern Pacific Ocean into the Arctic Ocean. The northern freshwater input from the atmosphere should increase as discussed above and salinity

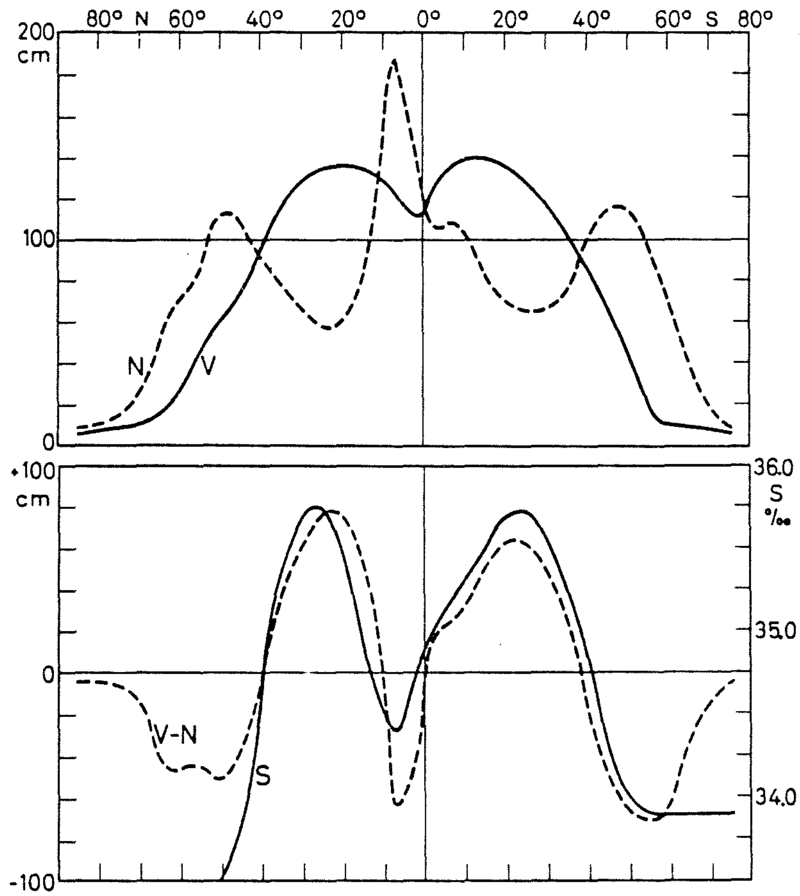


Fig. 3.4. Meridional distribution of annual averages of global zonal-mean precipitation P and evaporation E (above) as well as E-P and salinity S (below) at the surface of the world ocean, including adjacent seas (Figure taken from Dietrich et al., 1980).

of the Bering Strait contribution might decrease, owing to its northern origin in the Pacific Ocean. Any changes in the amount of water flowing through Bering Strait in an altered ocean circulation scenario are more difficult to characterize, and quantitative estimates of the expected changes in surface salinity are not easily made.

In summary, for the east coast of Canada north of approximately 45°N we expect that the initial (decadal) response to CO_2-induced climate changes will be warmer mixed layers and reduced salinities with geographical variations in the salinity changes similar to those in (P-E). Because of the higher mixed layer temperatures and lower salinities, the vertical stability of the near surface region should be enhanced, and combined with reduced wind stress, this should result in somewhat thinner mixed layers. However, in presently ice-covered regions which become ice-free (see Section 4), increased mixing by the surface wind stress will probably result in deeper mixed layers.

The changes discussed above should dominate the mixed layer's response to altered climatic conditions and changes should be realized on a decadal or shorter time scale. On this time scale, the second region which lies below the mixed layer of the "subtropical gyre" (the mid-latitude circulation which is closed on the western side of the North Atlantic by the Gulf Stream) but above the base of the permanent thermocline would also be influenced. Recent efforts to model this region are typified by the "ventilated thermocline" model of Luyten et al. (1983a). The basic premises of this model are the classical ideas that the region is renewed by water pumped downwards out of the mixed layer by convergences in wind-driven currents, and that once subducted away from the surface the fluid is inviscid, immiscible and conserves potential vorticity (i.e. total angular momemtum). As noted by Luyten <u>et al.</u> (1983b), "the main climatic inference [from their model] for tracers injected into a thermocline in steady state is that there are two time-scales in the subtropical thermocline: an advective time scale [≈ 30 yr] associated with distance from regions of direct ventilation of a density layer at the surface, and a subsurface diffusive time scale [≈ 300 yr] from ventilated to unventilated regions." In fact, Cox and Bryan (1984) suggest that a large "pool" of water located in the core of the subtropical gyre which the idealized model of Luyten <u>et al.</u> (1983a,b) predicts to be unventilated by vertical pumping, is strongly ventilated through lateral mixing with the intense western boundary current. Hence the time scale for renewal of this water is not greatly different from that of the region which is directly ventilated by vertical pumping (in agreement with Sarmiento's (1983) tritium measurements). On the other hand, the region which lies equatorward of the directly ventilated region has been shown to have a distinctly greater ventilation age than that of the rest of the gyre (Broecker <u>et al.</u>, 1978), and hence should not be strongly influenced by CO_2-induced climate change on decadal time scales.

Luyten <u>et al.</u> (1983b) have presented examples to illustrate that the directly ventilated region is less sensitive than expected to changes in Ekman pumping, but that its density stratification, mean flow and boundaries with the unventilated regions are all sensitive to changes in the surface density field which is determined within the mixed layer region. Thus in addition to the mixed layer, the mid- to low-latitude region above the base of the main thermocline is also sensitive to CO_2-induced climatic change on decadal time-scales, probably with tendencies for temperature and salinity changes similar to those of the mixed layer (i.e. tendencies to become somewhat warmer and saltier south of about 45°N).

At higher latitudes, the mid-depth waters are renewed by convective processes akin to those which maintain the mixed layer but extending to much greater depths. It should be noted that changes in this process may have a significant impact on climate. For example, Broecker et al. (1985) estimate that deep-water formation presently results in an estimated 5 x 10^{21} cal of heat being released to the atmosphere at high latitudes each year: 30% of the solar heating reaching the surface of the Atlantic Ocean north of 35°N. If deep-water convection were turned off, the heat loss to the atmosphere could be significantly reduced and large regional climate changes could be expected. The influence on climate predictions of changes in this heat transport mechanism is only recently beginning to be examined (e.g. Spelman and Manabe, 1984).

Most water formed by deep convection sinks to an intermediate level and then spreads horizontally along surfaces of constant potential density. Labrador Sea Water (2.9-3.4°C, 34.84-34.89 ppt) is a major water mass formed in this way in the western Labrador Sea (Clarke and Gascard, 1983). It fills much of the mid-depth (500-2000 m) northern North Atlantic Ocean and, according to Talley and McCartney (1982), is renewed on a decadal time scale. Thus, this water will also feel the influence of CO_2-induced climatic change within the first few decades. Since the renewal rate of this water is strongly affected by the surface salinity, the stratification of the surface waters and the action of the wind stress over the entire Labrador Sea (which acts to "dome up" the isopycnals there; Clarke and Gascard, 1983), it seems possible that initially (on a time-scale of decades) increased fresh water input at high latitudes, warmer surface layers and decreased winds would act in concert to reduce the renewal rate of Labrador Sea Water. Because of reduced renewal rates and hence isolation from the cold, fresh water input from the surface, we expect that under these conditions the salinity and temperature of Labrador Sea Water (or, more precisely, the water in the region presently occupied by Labrador Sea Water) will increase as a result of mixing with subsurface water masses of similar density. Note that this tendency is opposite to that expected in the overlying surface layer.

The abyssal water of the North Atlantic (the North Atlantic Deep and Bottom Waters which are denser than, and flow beneath the Labrador Sea Water discussed above) are formed by deep convection in the Norwegian and Greenland Seas. One might expect that the <u>initial</u> (decadal) response to climate change of the formation process of this water would be similar to that of Labrador Sea water. However, there are some important differences. First, the increase in fresh water input in these areas is unlikely to be as large as that in the Labrador Sea so a stabilizing halocline may not form. Second, there should be greatly reduced ice cover in this region resulting in more efficient mixing and extraction of heat from the sea surface. Although both these influences argue against reduced formation rates of Deep and Bottom Waters, the higher air temperatures and reduced wind stress in this region may still lead to lower formation rates as an initial response to expected climatic changes. Although it is true that enhanced stability is expected at high latitudes during summer, the small changes in predicted winter surface temperature using a simple mixed-layer ocean model (Fig. 3.2) suggest that the balance could go either way.

Because of the volume of the abyssal waters and their limited formation rate, it would take many decades (probably several centuries) for this region to react to climate change. On these very long time scales other factors need to be considered. A critical factor in abyssal water renewal is the relatively high salinity of the North Atlantic. If this characteristic were changed, deep water renewal and circulation could change drastically. Gordon and Piola (1983) summarize the arguments for why the North Atlantic is a relatively salty ocean. Basically, evaporation exceeds precipitation when averaged over the North and South Atlantic Oceans so there is a tendency for a continual increase in salinity. This tendency is at present balanced by an advective input of lower salinity water from other oceans to establish a dynamic equilibrium. If either the value of (E-P) averaged over the entire Atlantic increases or if the exchange rate with other oceans decreases (we do not know what controls the magnitudes of this exchange rate), the average salinity of the North Atlantic should increase further. Thus, <u>over the long term</u> (centuries), the influence of an enhanced atmospheric hydrological cycle, may be to increase the average salinity of the North Atlantic Ocean. Combined with more efficient mixing by surface wind stress due to reduced ice cover, this could result in enhanced deep convection. Indeed, Broecker <u>et al.</u> (1985) assimilate evidence from faunal, chemical and isotope studies to argue that deep-water production in the North Atlantic was reduced during glacial times, and hence one might expect enhanced deep-water formation to accompany higher levels of atmospheric CO_2. The abyssal waters in this new regime would eventually become considerably warmer than the present case, especially since they would derive their temperature characteristics from wintertime surface waters at high latitudes which show the largest temperature increase in the increased CO_2 scenarios (Bryan <u>et al.</u>, 1984). Further research is required before we can make conclusive statements on the fate of deep convection under the influence of CO_2-induced climate changes.

4. ICE AND SEA LEVEL RESPONSES

Ice plays a major role in the determination of climate change. When it is included in numerical models, an important feedback mechanism occurs. An increase in atmospheric CO_2 results in a warmer atmosphere and in response to the warming, the sea-ice extent and thickness are reduced (Fig. 4.1). This reduces the albedo (see equation (1): the albedo of ice is about 0.7 compared with a value of 0.3 for a global average) in these regions resulting in further local and global warming. This dependence on sea ice extent strongly influences the seasonal and geographical variations in temperature change seen in Figure 2.2. Models which neglect oceanic heat transport tend to overestimate the snow/ice albedo feedback effect. For example, in the 1 x CO_2 case shown in Figure 4.1 the seasonal signal in ice extent is somewhat too strong for this reason. Spelman and Manabe (1984) note that, in their idealized model, the poleward transport of heat by ocean currents [which is not included in Manabe and Stouffer's (1980) model] "raises the surface temperature at high latitudes, shifts poleward the margins of snow and sea ice, decreases the contribution of the albedo feedback effect (since both insolation and length of latitude circles decrease toward the poles), and reduces the sensitivity of the climate". Nevertheless, the inclusion of ocean heat transport does not change the qualitative conclusions discussed above.

Ice extent is, of course, important in its own right. An example of the influence of ice extent changes associated with climate changes of the order of those being predicted for a 2 x CO_2 scenario is provided by observations from the period 1880 to 1950 when the mean temperature of the Northern Hemisphere rose by about 0.6°C and, in the Arctic and northern Scandinavia, the mean temperatures in January and July changed by as much

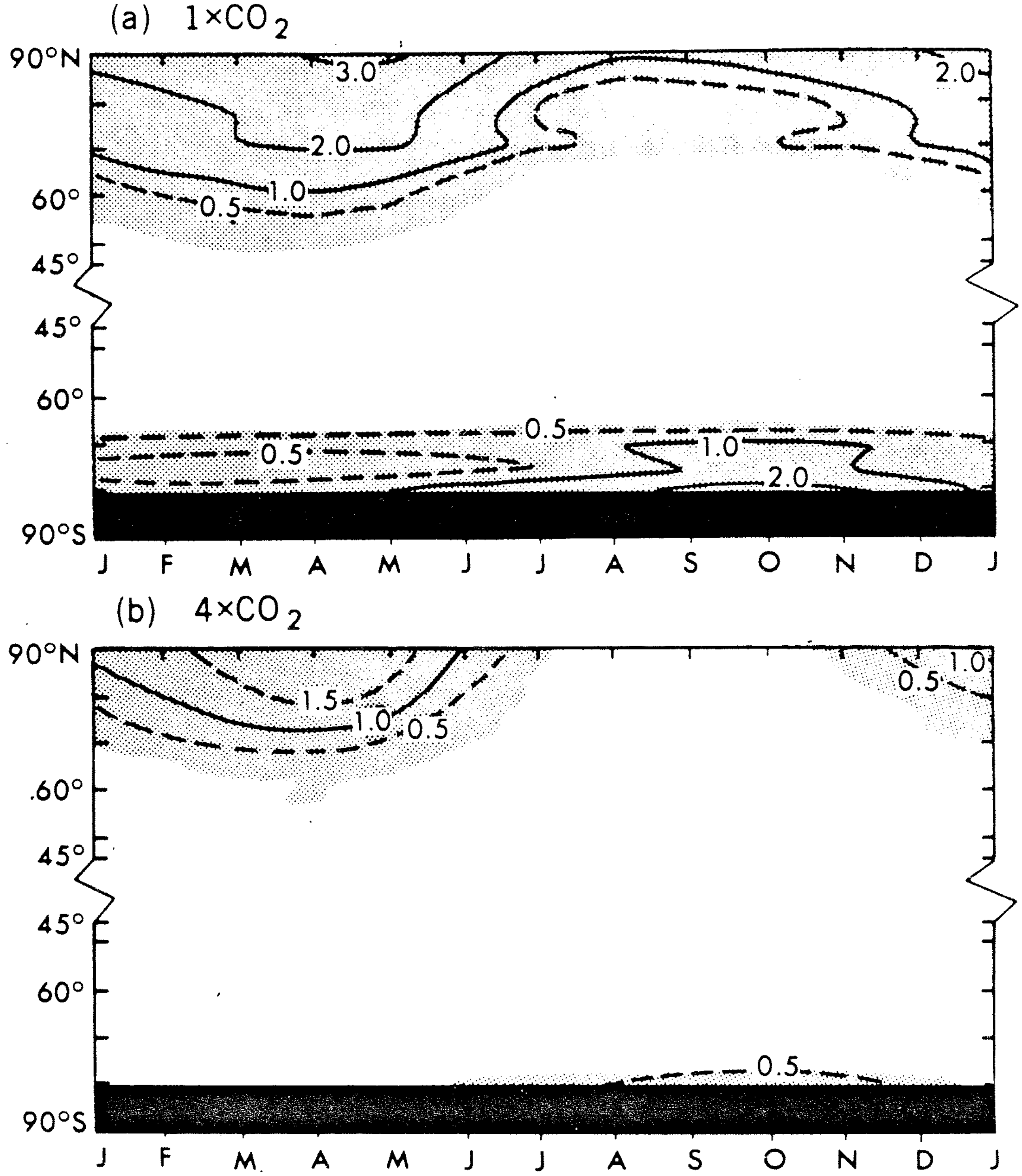


Fig. 4.1. Model predictions of the latitude-time distribution of sea ice thickness (meters): (a) 1 x CO_2 experiment; and (b) 4 x CO_2 experiment. Shading indicates regions where ice thickness exceeds 0.1 m. Changes in ice extent and ice thickness for the 2 x CO_2 case would be roughly half of those for the 4 x CO_2 case. (Figure taken from Manabe and Stouffer, 1980.)

as 6°C and 3°C respectively, from the coldest to the warmest decade (Wallen, 1984; Fig. 4.2). Ahlmann (1948) noted that in the Barents and Kara Seas ice conditions rapidly improved in the early 1920's and, by the end of the 1930's, the north coasts of Europe and Asia were practically ice-free from the end of August to the end of September, allowing sea transport through the Northeast Passage without the aid of icebreakers. Improved drift-ice conditions also resulted in the shipping season on West Spitzbergen increasing from three months around 1900 to seven months in the early 1940's.

These observations suggest that the change in ice extent due to climate change associated with doubled atmospheric CO_2 may have significant positive influences on Arctic sea transportation and exploration. Indeed, Parkinson and Kellogg (1979) estimate that for a 5°C increase in air temperature, the Arctic pack ice would be absent for August and September but would re-form in autumn. Hengeveld (1984) notes that studies of past climates and recent numerical model results suggest "a major retreat in southern ice limits, possible complete disappearance of ice in the Arctic ocean during mid-summer, and a Hudson Bay free of ice year-round."

There may also be significant changes in glacial ice with increased atmospheric CO_2 concentrations. The effects of atmospheric temperature change on the Greenland ice cap were investigated by Ambach (1980). For a temperature rise in northern latitudes of about 6°C [similar to the maximum magnitude of changes predicted by Manabe and Stouffer (1980) for a 2 x CO_2 scenario] an incremental rate of rise of sea level of about 24 cm per century was predicted, associated with melting of the Greenland ice cap. Assuming a doubling time for atmospheric CO_2 of order 100 yr, the average increment in the rate of rise between now and the time of doubled CO_2 would

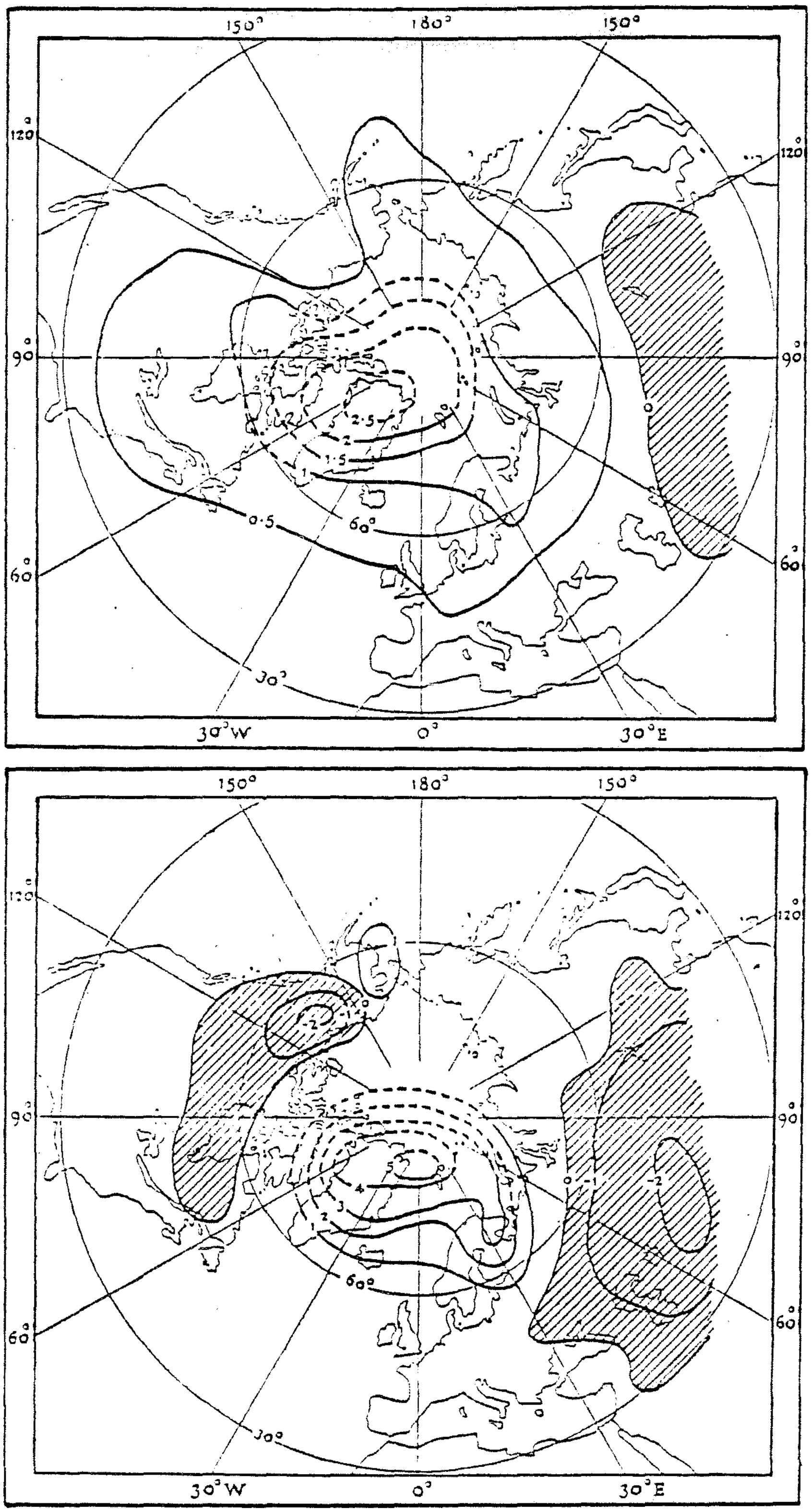


Fig. 4.2. Deviations of the mean annual (upper panel) and December (lower panel) temperatures in the Arctic in 1929-38 from the average for the period 1881-1938 (after E. Rubinstein).

be about 12 cm per century, corresponding to a decrease in ice volume equivalent to 480 km^3 per year. Since the Greenland ice cap is land based, much of this decrease will be due to melting but it is also likely that a warmer climate would be accompanied by an increased calving rate of icebergs from the Greenland ice cap. Even if the ice were not replaced by precipitation this could persist for many centuries and it is possible that increased snowfall over Greenland could result in a permanent increase in iceberg calving around Greenland.

The rise in sea level is also of concern. In addition to the possible sea level rise due to the melting of the Greenland ice-cap, Barnett (1983) estimates that a 20% reduction in the volume of alpine glaciers over 100 years would result in an additional rise of about 12 cm per century. Adding these increments to the observed rate of increase in sea level over the past 50 years of order 17 cm per century, Ambach (1980) estimates a probable rise in sea level during the next century of order 41 cm.

Although the rise in sea level estimated above would be significant for some coastal areas, the major concern of some investigators is that the rise in temperature associated with CO_2 increases may affect the West Antarctic ice sheet. While the Greenland and East Antarctic ice caps are land based and hence relatively stable, the West Antarctic ice cap is largely grounded below sea level and is considered unstable. Hengeveld (1983) states that a global warming of 3°C may be sufficient to remove the ice shelves reinforcing this ice sheet, allowing the entire ice sheet to transfer into the ocean rather rapidly. Some investigators have suggested that the West Antarctic ice sheet could melt in less than 100 years (e.g. Mercer, 1978; Thomas <u>et al.</u>, 1979) resulting in a rapid increase in sea level of order 5 m and an estimated reduction in continental surfaces of order 17%. On the other hand, the President's Science Advisory Committee in the U.S.A. reported that 400-4000 years might be required to melt the West Antarctic ice sheet, and recent numerical model results tend to support this more conservative scenario (Van dr Veen, 1985). Hansen et al. (1981) suggest that if the air above the ice sheet remains below freezing, the warmer temperatures could actually result in increased snowfall and net ice sheet growth which would lower the sea level. As Hengeveld (1984) notes, the latter possibility could result in increased glacial flow and hence increased iceberg production.

From the above it is apparent that ice extent over the Arctic Ocean and adjacent regions is likely to decrease with increased atmospheric CO_2 concentration, but the changes in sea level are strongly dependent on what happens to the Antarctic ice cap - a question which remains unanswered.

5. LARGE-SCALE WIND-DRIVEN OCEAN CIRCULATION

The major features of the present-day mean atmospheric surface circulation over the North Atlantic Ocean are the easterly Trade Winds to the south and the prevailing Westerlies in temperate latitudes. This pattern gives rise to a maximum westward wind stress on the ocean surface in a latitude band centered near 15°N and a maximum eastward wind stress in a band extending northeastward from Cape Hatteras to the Faeroe Islands (Leetma and Bunker, 1977). The north-south component of North Atlantic wind stress is relatively small in general except for limited areas south of Greenland and off the west coast of Africa near 20°N.

The direct action of wind stress on the ocean gives rise to a near-surface water movement (Ekman transport) with a net transport perpendicular to the direction of wind stress because of the Earth's rotation (Veronis, 1980). The Ekman transport is directed northward south of about 30°N and southward north of this latitude in the North Atlantic. It is significant in the global transport of heat by the oceans because it involves meridional displacements of relatively warm surface waters (Bryan, 1982). Bryden and Hall (1980) estimated that the near-surface Ekman transport in the North Atlantic at 24°N contributes about 40% of the net meridional heat transport across that latitude, but provides only about 17% of the mass transport carried by the geostrophic gyral circulation.

The largest part of the wind-driven ocean circulation is associated with horizontal divergences and convergences of the surface Ekman transport. These are proportional to the spatial curl (representative of the effective torque) of the horizontal wind stress, and so depend on both the magnitude and the spatial patterns of the atmospheric winds. The horizontal geostrophic flow in the interior of the ocean adjusts to balance the wind-driven Ekman divergences and convergences, communicating with the near-surface Ekman layers through vertical velocities set up at their base.

The spatial patterns of surface winds over the North Atlantic give rise to several large horizontal cells or gyres which constitute the major portion of the wind-driven circulation. As an illustration, Figure 5.1a shows the transport streamlines predicted by a numerical model of the North Atlantic circulation (Anderson _et al._, 1979) with forcing by an approximation (Fig. 5.1b) to the observed mean zonal wind stress. The model includes realistic topography and density stratification, but neglects nonlinear effects. Many of the features in this idealized model have their counterparts in the observed circulation of the North Atlantic Ocean.

The mean wind stress gives a negative (clockwise) wind stress curl in the latitude band between approximately 15°N and 45°N, resulting in the

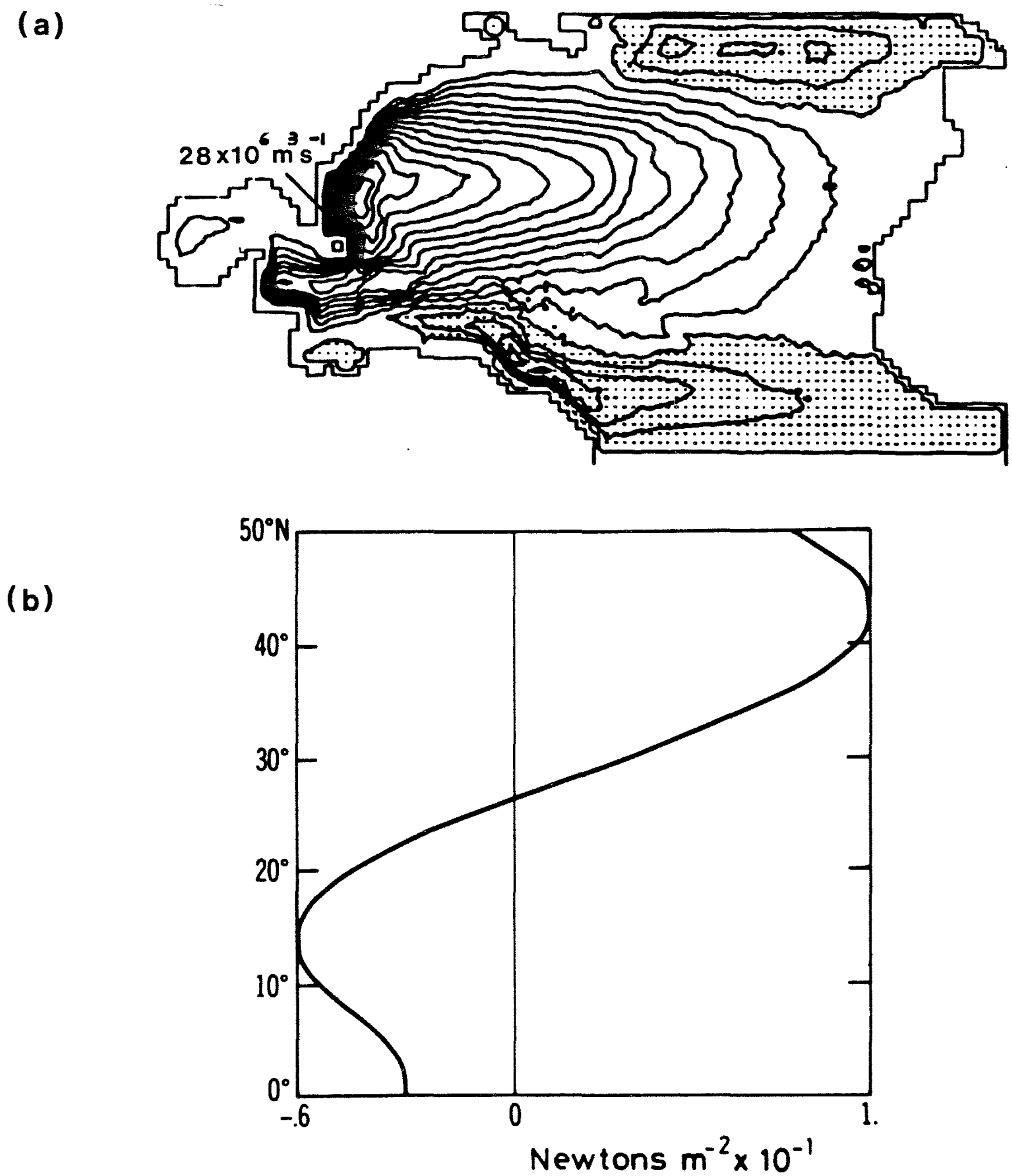


Fig. 5.1. Idealized linear, stratified, wind-driven circulation of the North Atlantic (Anderson et al., 1979): (a) stream function with a contour interval of 2 x 10^6 m^3 s^{-1}; (b) plot of the westward wind stress as a function of latitude used in the model. This has extrema at 14°N and 43°N, a zero crossing at 27°N, and zero curvature at 8°N and 29°N.

large mid-latitude gyre in Figure 5.1a. The negative wind stress curl produces a horizontal convergence of water in the surface Ekman layer, a downward vertical velocity at the base of the Ekman layer, and a southward geostrophic flow in the interior. The total of this southward interior flow plus any southward Ekman transport (together constituting the Sverdrup transport) is balanced by a northward return flow in a western boundary current which, in the North Atlantic Ocean, we know as the Gulf Stream. In the simplest wind-driven models the transport in the boundary current is directly proportional to the zonal average curl of the wind stress, and for a fixed spatial pattern of wind stress, is directly proportional to the magnitude of the stress. Observations indicate that the mean transport in the Florida Current at the upstream end of the Gulf Stream nearly matches the theoretically predicted values for linear wind-driven flow with realistic wind forcing (Leetma _et al._, 1977).

Simple models, such as that shown in Figure 5.1a, do not predict the observed fact that the Gulf Stream gains in transport on moving downstream, and also veers away from the coast near Cape Hatteras at about 35°N to flow eastward as a mid-ocean jet. At its maximum south of Nova Scotia, the Stream carries on the order of five times the predicted wind-driven transport (Worthington, 1976). Models of wind-driven circulation that include nonlinear dynamics can reproduce some of the features of the observed Gulf Stream intensification. The downstream increase in transport is partly associated with recirculation in a relatively narrow gyre offshore of the Stream which is embedded in the larger wind-driven gyre (Veronis, 1980).

The observed separation of the Gulf Stream from the coast is a significant feature of the North Atlantic circulation, but the dynamics that control the location of the separation point and the subsequent path of the current in mid-ocean are not completely understood. In the simplest wind-driven models, the mid-ocean Gulf Stream analogue follows the contour of zero wind stress curl, a natural internal boundary since the linear model allows no net transport of water across this contour in the interior of the ocean. In fact, the zero contour of wind stress curl in the temperate North Atlantic leaves the coast of North America near 35°N at approximately the observed Gulf Stream separation point, and the zero contour and the mean position of the Gulf Stream are nearly coincident across the western North Atlantic (Fofonoff, 1980).

A second mechanism for the separation of a wind-driven western boundary current from the coast has been proposed by Parsons (1969) and studied further by Veronis (1973) and Huang (1984). In a two-layer wind-driven model of the North Atlantic subtropical gyre, the southward geostrophic transport in the upper layer interior is proportional to the westward increase in the square of the thickness of the upper layer from the eastern boundary to the offshore edge of the western boundary current, and the northward geostrophic transport in the western boundary current is proportional to the decrease in the square of the thickness of the upper layer across the current. Since the boundary current transport is greater than the interior geostrophic transport by the amount of the southward Ekman transport, the minimum thickness of the upper layer occurs at the inshore edge of the boundary current. With a specified spatial distribution of wind stress, there is a critical strength of the winds and resulting Ekman transport which causes the upper layer depth to go to zero there. This defines the separation point, and in the model the boundary current leaves the coast and flows northeastward in the interior, marked by the intersection of the lower layer and the ocean surface. In this model, variations in the position of the mid-ocean Gulf Stream can be caused by increases or decreases in wind stress without any changes in spatial wind patterns, with stronger (weaker) winds causing a southward (northward) shift in the separation point and the internal boundary. Figure 5.2 from Parsons (1969) shows these changes in Gulf Stream position as a function of a dimensionless wind stress parameter. The northern boundary in the model domain for Figure 5.2 coincides with the zero wind stress contour. In a more elaborate version, Veronis (1973) extended the model domain northward so that the zero wind stress curl line was an internal boundary. The zero curl contour remains a major dynamical constraint providing the ultimate northern boundary for the subtropical gyre (according to linear theory, at least), but the Gulf Stream separates from the western boundary before reaching the latitude of zero wind stress curl because of the mechanism discussed above.

The separation point and deep-ocean trajectory of the Gulf Stream may also be influenced by bottom topography and coastline geometry, independent of the details of forcing. Recent evidence from Atlantic Oceanographic Laboratory measurements suggests that both the Gulf Stream and the North Atlantic Current extend to the seafloor. As a result, there should be a dynamical tendency for the flow to proceed along geostrophic contours defined by constant values of Coriolis parameter divided by the water depth (f/H) (Warren, 1963). On the other hand, the Gulf Stream leaves the coast at Cape Hatteras without any marked change in direction, the coastline itself veering sharply to the left of the stream's path. The separation may simply be related to inertia and the changing shape of the coastline (Fofonoff, 1980).

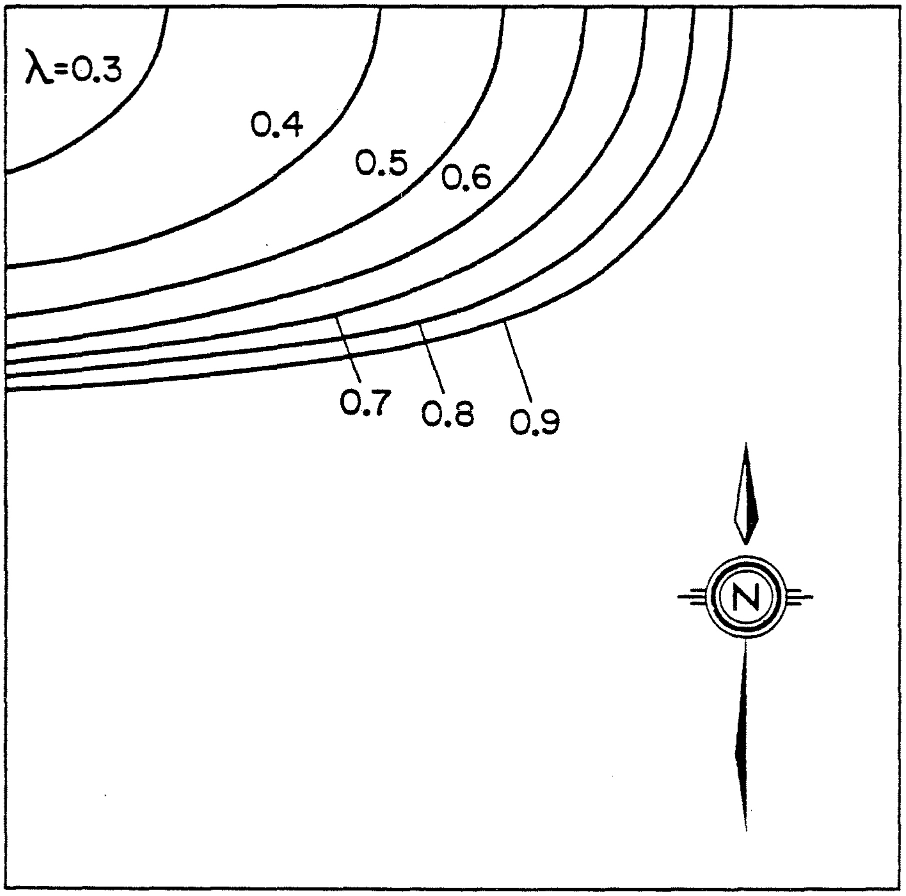


Fig. 5.2. Separation point and subsequent path of the Gulf Stream as a function of wind stress in a model ocean: larger wind stress results in separation at a more southerly latitude (Parsons, 1969).

The Gulf Stream is subject to energetic time-dependent meanders, some of which detach themselves from the body of the current and form isolated current rings (Richardson, 1983). The formation processes are as yet uncertain, but the primary energy source for these fluid dynamical instabilities lies in the strong flow and potential energy associated with the mean Stream. Numerical simulations of wind-driven flow with high horizontal resolution have shown such instabilities, and as might be expected, increased (decreased) mean wind forcing results in a more (less) energetic time-dependent circulation (Holland and Lin, 1975). The process is certainly nonlinear, and the Reynolds stresses associated with the eddies may play an important role in the dynamics of the mean Gulf Stream and its recirculation (Holland, 1978).

Gulf Stream eddies are an important means of horizontal exchange of momentum and water properties between the Gulf Stream and Sargasso Sea, on the one hand, and the Slope Water found north of the Stream between Cape Hatteras and the Grand Banks, on the other. Once introduced into the Slope Water, warm-core Gulf Stream rings also provide a potentially important mechanism for exchanging water between the Coastal and Slope Waters; the rings have been observed to entrain surface water from the offshore edge of the shelf and carry it seaward into the Slope Water, and Smith (1978) suggests that some of the lost surface water is replaced by deep flows of Slope Water onto the Scotian Shelf. Approximately five warm core rings are formed each year. The rings drift slowly westward in the Slope Water, and eventually are re-absorbed by the Stream after a mean lifetime of about six months (Richardson, 1983). The currents in detached Gulf Stream rings are representative of actual Gulf Stream flows at least in their initial stage, so a stronger (weaker) Gulf Stream would tend to produce more (less) energetic eddies. The dynamics controlling the frequency of eddy formation events are not known, but intuition suggests that the number of eddies produced in a given time should also be generally proportional to the energy in the mean Gulf Stream.

As discussed earlier, after leaving the coast, the Gulf Stream flows roughly along the zero wind stress curl contour. In the present-day northern North Atlantic this contour trends towards the northeast with a central position near 45°N. North of this contour the wind stress curl is positive and the result is a counterclockwise gyre which is reproduced in the model calculation in Figure 5.1a. The northeastward flowing current separating the northern gyre from the subtropical Gulf Stream gyre is called the North Atlantic Current. This current is generally considered to include a component of Gulf Stream transport, augmented by Slope Water originating north of the Stream. The North Atlantic Current is associated with a Sub-Polar Front with a near-surface decrease in temperature of order 5°C on moving northward across the front (Clarke, 1984). Thus, changes in the path of the North Atlantic Current can produce significant temporal changes in the near-surface water temperature in the northern North Atlantic. Possible dynamical constraints on the path of the North Atlantic Current include the position of the zero wind stress curl contour, the strength of the westerly winds for a fixed spatial pattern, and the guiding effects of bottom topography just as discussed for the Gulf Stream.

There is some observational evidence linking decreased sea surface temperature in the North Atlantic north of about 50°N with stronger westerly winds (Lamb, 1972; Colbrooke and Taylor, 1979). Figure 5.3 from Lamb (1972) shows the difference in surface temperature in the North Atlantic between early nineteenth century conditions and those of a later reference

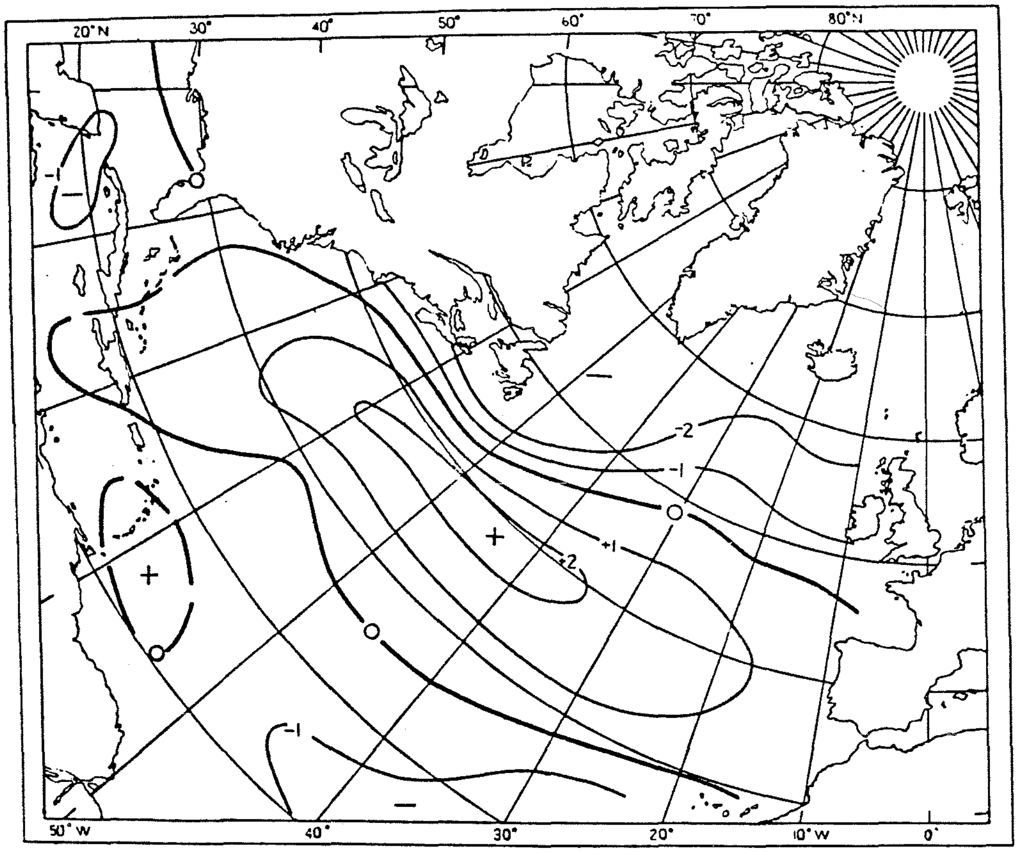


Fig. 5.3. Departures of average surface water temperatures (°C) in the first half of the nineteenth century from the values for the periods 1887-99 and 1921-38, using Maury's collection of the observations from American ships (65% of the observations made in 1840-8, 18% 1830-9, 10% 1820-9, 7% between 1780 and 1819). (Figure taken from Lamb, 1972).

period centered early in the twentieth century. The sea surface in the northernmost North Atlantic was up to 2°C cooler during the earlier period, while at the same time warmer conditions prevailed in a region centered near 40°N. Variations on Parsons' (1969) dynamical model of Gulf Stream separation from the coast have been invoked to explain these observations: stronger winds increase the water transport in the interior of the subtropical gyre and in the return flows, the thermocline deepens to maintain a geostrophic balance in the upper layers of the circulation, and for a fixed volume of upper layer water, the area of the gyre decreases and the boundary currents shift to the south. The northern cooling is then due to the southward shift of the sea surface temperature gradients of the Sub-Polar Front. Figure 5.4 shows the mean March sea surface temperature of the northwestern North Atlantic (NOAA, 1982) with some of the temperature anomaly contours from Figure 5.3 superimposed. It is apparent that 2°C changes can be explained by spatial shifts of the mean field by only a few hundred kilometers. The mid-latitude warming under increased wind forcing may be attributed to a greater eastward penetration of the warmest waters carried by the stronger Gulf Stream jet.

Some of the North Atlantic Current transport recirculates to the north and west in the northern, counterclockwise gyre, re-entering the Labrador Sea south of Greenland as the warm Irminger Current and contributing to a southward flow in the western Labrador Sea. This southward flow is another wind-driven western boundary current (analogous to the Gulf Stream), and as such its strength is proportional to the strength of the atmospheric forcing. The relatively warm and saline Irminger Current and its continuation around the Labrador Sea provide one of the source waters for Labrador Sea Water (Clarke and Gascard, 1983). The warmer waters also

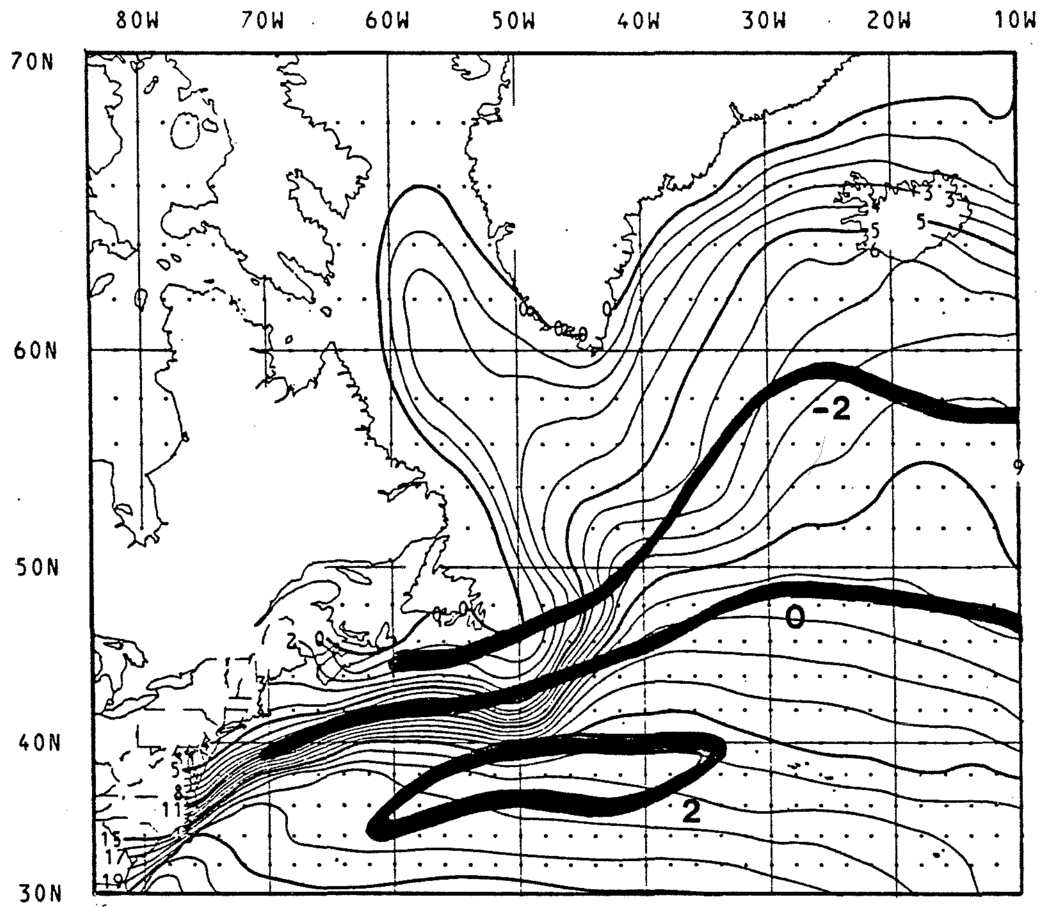


Fig. 5.4. Climatological mean sea surface temperature of the North Atlantic Ocean for the mouth of March (from NOAA, 1982) with some temperature anomaly contours from Figure 5.3 superimposed.

mix with cold, relatively fresh Arctic outflows to produce a characteristic range of temperature and salinity properties in waters carried by the Labrador Current. In the next section, the shallow Labrador Current is discussed as a primarily buoyancy-driven flow, but the warmer waters transported by the larger scale wind-driven circulation in the Labrador Sea influence the temperature-salinity characteristics of the Labrador Current through this mixing.

This section has concentrated on purely wind-driven circulation, but the oceans also have a thermohaline circulation driven by differential surface heating and the global distribution of precipitation and evaporation. The transport and trajectories of western boundary currents, such as the Gulf Stream, can also be influenced by the thermohaline circulation. For example, Holland (1972) discusses a numerical model applicable to the North Atlantic circulation that combines wind and thermal forcing, and the effects of variable bottom topography. The thermal forcing sets up a vertical circulation cell with sinking at high latitudes and a southward transport in a deep western boundary current. The combined effects of stratification and bottom topography result in the northward return flow in the thermally-driven cell being concentrated in an upper level western boundary current which adds significantly to the transport of the wind-driven westward boundary current in the model. Worthington (1976) discusses the possibility that the Gulf Stream intensifies after severe winters in response to the convective formation of a deeper mixed layer of 18°C water and the associated observed deepening of the thermocline beneath the mixed layer.

Based on the dynamical considerations and historical observations discussed above, we can propose a qualitative scenario for changes in the

North Atlantic Ocean circulation associated with a weakened atmospheric general circulation as postulated in Section 2.

In this scenario of weakened wind forcing, the Ekman transport and the divergences and convergences of the Ekman transport decrease in magnitude, the geostrophic transport in the ocean interior decreases in magnitude, and the Gulf Stream and the North Atlantic Current decrease in transport and in peak speed. In the Gulf Stream and its recirculation south of Nova Scotia, and in the North Atlantic Current, the decrease in energy of the mean flow is accompanied by a reduction in the frequency of occurrence and strength of time-dependent fluctuations of the flow in general and of Gulf Stream rings in particular. Furthermore, the wind-driven transport of warm, saline water into the Labrador Sea via the Irminger Current decreases.

In a more speculative vein, it is possible that the mean paths of the Gulf Stream and the North Atlantic Current could shift somewhat to the north under conditions of weaker winds for a fixed spatial wind pattern. The models of separated western boundary currents predict several distinct regimes, within each of which there is a sensitive non-linear relationship between current path and wind forcing (Huang, 1984). Any quantitative estimate of path shift for a specified change in wind stress depends explicitly on a whole family of parameters and is beyond the scope of the present discussion.

The hypothesized shifts in boundary current paths associated with relatively weaker winds in an increased CO_2 scenario should tend to alter results such as those shown in Figure 3.2 by contributing additional sea surface temperature changes with signs opposite to those in Figure 5.3. The mean surface warming and associated atmospheric changes predicted for a doubled CO_2 scenario are much larger than recent natural climate fluctuations, and we might propose an accompanying shift in the North Atlantic Current that would be large compared to recently observed variations. There are major uncertainties in such an approach: the ocean circulation also plays a part in the dynamics of the atmospheric circulation, notably through the controlling effects of ocean surface temperature on the fluxes of heat and moisture to the overlying atmosphere. For example, it has been suggested that the position of the zero wind stress curl contour could be partly controlled by the presence of the Gulf Stream, with the associated strong cross-stream temperature change (Fofonoff, 1980). The validity of such suggestions remains to be determined. Further understanding will come with the development of fully-coupled ocean-atmosphere models.

A decrease in the strength of the Gulf Stream and its eddies should decrease the lateral transfer of momentum and water properties from the Stream and the subtropical gyre to the Slope Water lying to the north. The temperature-salinity properties of the Slope Water are determined in a general sense by the temperature-salinity properties and relative amounts of warm waters of Sargasso Sea origin and cold waters of Labrador Current origin, which together form the bulk of the Slope Water (McLellan, 1957). In the next section, it is suggested that the Labrador Current input to the Slope Water might increase in a climate regime with increased high-latitude precipitation, such as proposed in current atmospheric scenarios for increased CO_2. Combined with the expected warming of both Sargasso Sea and Labrador Current waters associated with the global change, the net effect on the Slope Water should be an increase in its absolute temperature but a decrease in its temperature relative to the new global average. The change in salinity of the Slope Water is less clear because of the competing tendencies toward reduced salinity due to the increased input from the fresher Labrador Current and toward increased salinity due to the higher salinity of its largest constituent, Sargasso Sea water. But the relative change should be fresher Slope Water compared to the new Sargasso Sea water.

A decrease in the number and intensity of warm-core rings in the Slope Water should also tend to decrease the overall levels of current variability and horizontal mixing therein. Such a decrease should reduce the offshore flux of waters from the outer continental shelf due to entrainment by Gulf Stream rings, and reduce the exchange of properties between the Slope Water and the continental shelf waters.

6. THE CONTINENTAL SHELF REGIONS

6.1 Our Understanding of the Present Regime

The physical oceanography of the various coastal and continental shelf waters off eastern Canada is dominated by the generally along-coast flow of the Labrador Current and the St. Lawrence River run-off system augmented by coastal run-off (Fig. 6.1; also see Fig. 3.3), and by the seasonal modification of water mass properties due to the exchange of heat, moisture and momentum with the overlying atmosphere. The major physical features of these waters, in the context of ocean climate and marine fisheries, have been summarized on a regional basis in a recent Department of Fisheries and Oceans report (Akenhead et al., 1981). That description will be taken as the starting point for the present discussion of what determines the physical oceanographic properties of these waters and of the changes that may be associated with a CO_2-induced change in the atmosphere-ocean-cryosphere system.

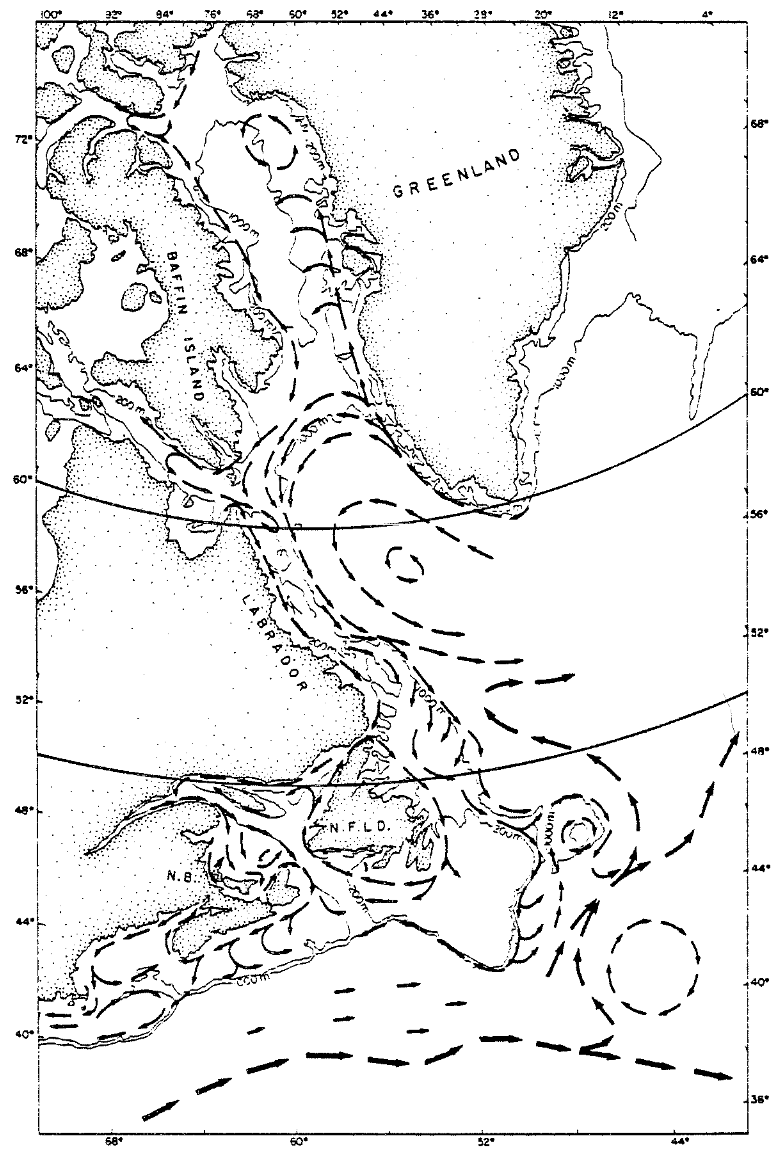


Fig. 6.1. Major surface currents off the east coast of Canada (adapted version of figure compiled by R. Reiniger and C.R. Mann).

We shall consider the following shelf regions (Fig. 6.2): the Labrador Shelf (including the Northeast Newfoundland Shelf south to 49°N), the Grand Banks of Newfoundland, Flemish Cap, the Gulf of St. Lawrence, the Scotian Shelf, and the Gulf of Maine region.

Throughout these regions, the vertical structure of the water column undergoes a pronounced seasonal cycle. During winter, atmospheric cooling and stirring result in the formation of relatively deep, surface mixed layers with thicknesses of order 100 m. These layers extend to the seafloor in many shallow areas, and elsewhere they usually overlie a denser bottom layer of relatively warm and salty water of offshore origin (e.g. water of Slope Water origin on the Scotian Shelf and in the Gulf of Maine). As a result of solar heating and freshwater input from local run-off, advection and melting ice, a seasonal pycnocline develops in the upper water column in spring and summer in most areas. The thickness of the surface mixed layer is then reduced to the order of 10 m and there is an intermediate layer of remnant winter water below the pycnocline and overlying the bottom water, resulting in a mid-depth temperature minimum. The thicknesses, property values and seasonal extents of these layers vary from region to region depending on the dominant local processes.

Over the Labrador Shelf, the southward-flowing Labrador Current has a major influence on the water mass properties. Although it is generally considered that some of the southward flow in the western Labrador Sea is part of the wind-driven counterclockwise gyre in the northern North Atlantic (Section 5), recent observations (Lazier et al., 1985) indicate that there is a seasonal increase in the Labrador Current's transport associated with the seasonal salinity minimum and run-off maximum. This suggests that the Current is driven primarily by buoyancy forces associated

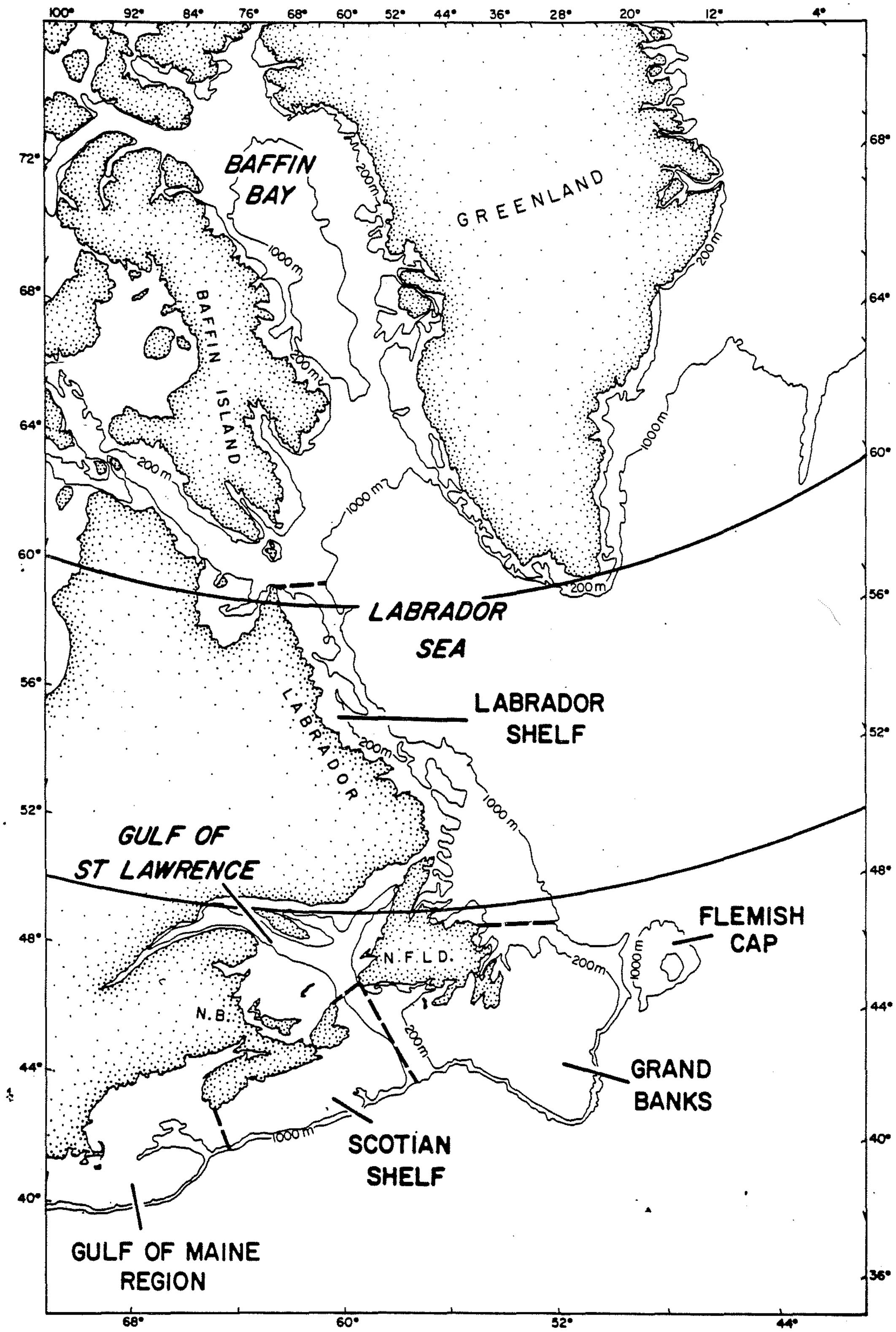


Fig. 6.2. Map showing the approximate extent of the major oceanographic regions referred to in Section 6.

with fresh water derived from precipitation over the polar region and the Hudson Bay drainage area, and from the seasonal melting of Arctic ice. Wind forcing may, however, modify the current through both local influences and effects on the strength of upstream flows such as the East Greenland Current, and there may be some interaction between the Current and the wind-driven southward flow in the western Labrador Sea.

The main branch of the Labrador Current flows along the continental slope and is associated with a shelf-break front, but there is also a weaker inshore branch along much of the Shelf which is augmented by buoyancy forcing associated with run-off from the Labrador coast (Lazier et al., 1985). The stratification and water mass properties over the Labrador Shelf are thus strongly influenced by the advection of water and ice from the north and by local run-off, as well as by exchanges with the atmosphere and by vertical and horizontal mixing. Seasonal ice-cover over much of the region (Fig. 6.3) interrupts the exchange of heat and momentum with the atmosphere during part of the year. The dominant mechanisms of cross-shelf exchange are uncertain, but surface Ekman transport and internal frontal circulation (Ikeda, 1985) and eddies associated with instabilities in the Current (LeBlond, 1982) are candidate processes. The southern portion of the Labrador Shelf may also be influenced by extreme northward excursions of the North Atlantic Current.

The Labrador Current also strongly influences the physical oceanography of the Grand Banks and Flemish Cap. The inshore branch primarily flows around southeastern Newfoundland as the Avalon Channel Current, while the offshore branch splits near 48°N, one part continuing south through Flemish Pass and along the eastern edge of the Grand Bank before swinging (in part) westwards, and the other veering east and flowing north of

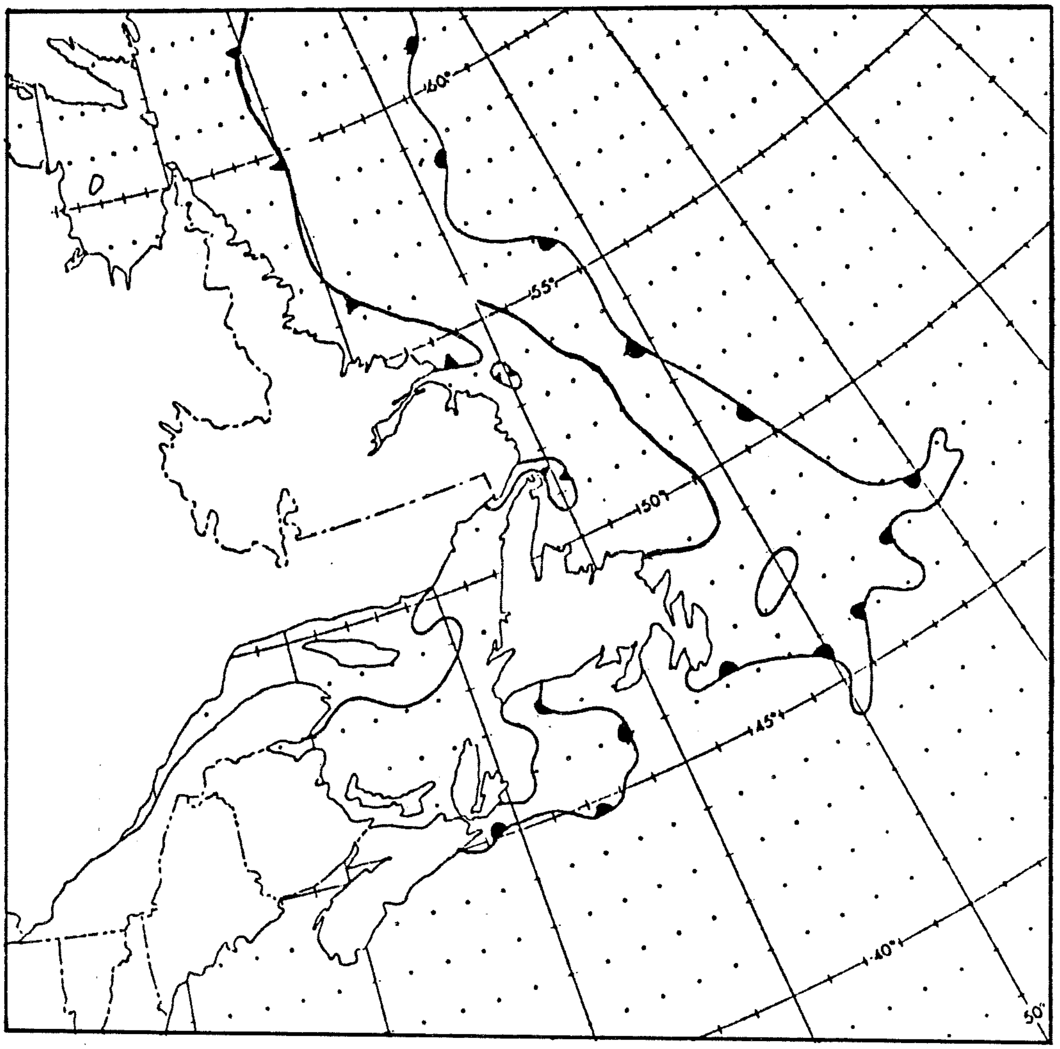


Fig. 6.3. Median, maximum and minimum extents of all types of sea ice (not iceberg) off eastern Canada for the first week of April. The median ice edge is based on a 10-year data base covering 1964 to 1973, and the minimum and maximum positions are for this period plus the following five years (from Sowden and Geddes, 1980).

Flemish Cap (Petrie and Anderson, 1983). The residual flow is thus primarily around the perimeter of these regions. The physical properties of the waters over the Grand Bank are influenced by the weak component of the Labrador Current (including ice, see Fig. 6.3) drifting across the Bank (Petrie and Anderson, 1983), local exchange with the atmosphere, and horizontal exchange associated with instabilities in the perimeter currents (Petrie and Isenor, 1985), wind-induced upwelling along the edge of the Bank (Kinsella, 1984), and episodic meanders and eddies from the Gulf Stream and North Atlantic Current to the south and east (Voorheis et al., 1973). Over the deeper Flemish Cap, water properties are strongly influenced by mixing products of Labrador Current and North Atlantic Current waters found south of the Cap (C.K. Ross, pers. comm., 1985).

The primary factors influencing the physical oceanography of the Gulf of St. Lawrence are the freshwater discharge from the St. Lawrence and other rivers (dependent primarily on precipitation over the Great Lakes and southern Quebec drainage areas), the inflow of water of Labrador Current origin (mainly through Cabot Strait and to a lesser extent through the Strait of Belle Isle), and heat and momentum exchange with the atmosphere. The oceanographic influences of freshwater discharge have been discussed in detail in another Department of Fisheries and Oceans study (Bugden et al., 1981). The circulation in the Gulf is in part estuarine in nature, with an outflow of relatively fresh water near the sea surface and an inflow of denser water through Cabot Strait, controlled primarily by the rate of freshwater discharge and the level of vertical mixing in the system. This mixing is of tidal origin in the St. Lawrence Estuary, but elsewhere in the Gulf the mixing is weaker and associated with local wind-forcing. There is considerable horizontal structure in the circulation.

The surface outflow forms a buoyancy-driven coastal current around the Gaspé Peninsula which broadens over the western half of the Gulf, and there is an inflow of surface water in the eastern Gulf. There is also a strong seasonality in circulation and stratification in the Gulf, with the spring/early-summer maximum in discharge resulting in a freshening of the southern Gulf, and in a salinity minimum and transport maximum in Cabot Strait that lag the discharge maximum by about four months (Sutcliffe et al., 1976). Under typical wintertime conditions, the Gulf is ice-covered from late January until mid-April (Sowden and Geddes, 1980).

Over the Scotian Shelf there is a general southwestward flow concentrated near the coast (Smith et al., 1978). The occurrence of a salinity minimum and transport maximum on the Halifax section several months after the peak discharge into the Gulf of St. Lawrence (Sutcliffe et al., 1976; Drinkwater et al., 1979) suggests that this flow is driven in part by the outflow from the Gulf through Cabot Strait. The physical oceanographic properties are also influenced by Coastal Water of Labrador Current origin (in part) flowing from the east (McLellan, 1957), and by offshore Slope Water (Section 5) as evidenced by the Slope Water at depth in some of the Shelf's basins. Stratification over the Shelf is controlled primarily by local air-sea heat exchange, the inflow of relatively fresh water from the northeast, vertical mixing associated with tidal and wind-driven currents, and cross-shelf exchange associated with wind-induced upwelling over the shelf-break (Petrie, 1983) and Gulf Stream eddies (Smith, 1978).

The outflow from the Gulf of St. Lawrence also influences the Gulf of Maine (Sutcliffe et al., 1976). Inflow from the Scotian Shelf combines with run-off from local rivers to drive a generally counterclockwise summertime circulation over its major basins (Bigelow, 1927). A second primary influence in the Gulf of Maine region is the strong M_2 tidal currents which drive year-round clockwise circulations around Georges and Browns Banks (Greenberg, 1983). The strong tidal currents also maintain year-round vertically well-mixed waters in many of the region's shallow areas with associated summertime tidal fronts around these areas (Garrett et al., 1978). Over the outer shelf, there is a continued southwestward flow of Coastal Water associated with the shelf-break front, whose surface signature is closer to the shelf-break here than to the east. The deep basins of the Gulf are supplied with Slope Water inflowing at depth through Northeast Channel (Brooks, 1985) and partially controlled by the wind stress over the Gulf (Ramp et al., 1985). Exchange with the offshore waters is also strongly influenced by Gulf Stream eddies impinging on the region.

6.2 Tendencies for Change

Some of the driving forces of the present shelf regime and the assumed signs (based on the available information and understanding as discussed earlier) of their changes associated with a doubling of atmospheric CO_2 concentrations are summarized in Table 1. The available information suggests that the magnitudes of the changes should generally be small, say of order 5 to 10 per cent. The spatial and seasonal variability of these changes in forcing will not be considered in detail, but the suggestions are that the increases in air temperature and precipitation should be larger both in winter and in northern latitudes. In particular, Figure 2.3 suggests that increases in precipitation minus evaporation are expected only at latitudes north of approximately 45°N so that the proposed

Table 1. Atmospheric and oceanic influences on the physical oceanography of eastern Canadian shelf regions, and the assumed sign of their changes associated with a doubling of CO_2 concentrations in the atmosphere. (The magnitude of the changes is expected to be of order 10% of present values.)

Driving Force	Assumed Sign of Change
Atmospheric:	
Air temperature	+ (4°C)
Precipitation	+
Precipitation minus Evaporation (north of 45°N)	+
Magnitude of large-scale wind stress curl	-
Average magnitude of wind stress	-
Oceanic:	
Gulf Stream/North Atlantic Current Position	No change except north and east of Flemish cap
Gulf Stream/North Atlantic Current Transport	-
Gulf Stream/North Atlantic Current Variability	-
Labrador Current Position	No change
Labrador Current Transport	+
Labrador Current Variability	+
Quantity of Arctic ice entering region	-

effects of increased freshwater discharge into the Gulf of St. Lawrence (for example) should be smaller and are more tentative than those suggested for more northern run-off. (In fact, Cohen (1985) suggests from detailed examination and interpretation of the predictions of two atmospheric numerical models that, with doubled atmospheric CO_2 concentrations, there will be a reduction in the net water outflow from the Great Lakes which accounts for about 30% of the total freshwater supply to the Gulf.) The assumed changes in the large-scale oceanic circulation are based on the premises that the Gulf Stream and North Atlantic Current are primarily wind-driven, the Labrador Current is primarily buoyancy-driven, and the position of these currents will not change significantly, except for a possible northward shift of the North Atlantic Current east and north of Flemish Cap.

The changes in the physical oceanographic features and properties of east coast shelf waters that seem most likely to be associated with these changes in driving force will now be discussed.

(i) Along-shelf advection and estuarine circulation

As a result of increased precipitation, run-off and melting of ice (the latter being part of the transient adjustment to warmer temperatures) in northern latitudes, both the inshore and continental slope branches of the Labrador Current should be stronger, providing increased transport along the Labrador Shelf, around and over the Grand Banks and Flemish Cap, and around southeastern Newfoundland. This change is consistent with the observations of a seasonal increase in current on the Labrador Shelf (Lazier *et al.*, 1985) and in Avalon Channel (C. Anderson, pers. comm., 1985). A stronger Labrador Current should also result in an increased flow of Coastal Water along the Scotian Shelf, an increased contribution of Labrador Current water to Slope Water, and on a more speculative level, increased inflows to the Gulf of St. Lawrence through Cabot Strait and the Strait of Belle Isle.

In view of the observed seasonal cycle in transport and salinity of the Cabot Strait outflow and consistent with the behaviour of a moderately stratified estuary such as the St. Lawrence Estuary where tidal mixing dominates, an increase in the overall estuarine circulation in the Gulf of St. Lawrence is expected to be associated with increased freshwater discharge into the Gulf. This should include increased surface outflow and return inflow at depth in the Estuary, a stronger Gaspé Current, and a further contribution to increased exchange through Cabot Strait. However, since parts of the Gulf are more similar to a highly stratified estuary whose response to increased freshwater discharge is reduced entrainment and transport (Bugden _et al._, 1981), the magnitude of the changes in transport may vary within the Gulf and the fractional increase in transport through Cabot Strait is likely to be less than that in discharge. This tendency towards increased circulation in the Gulf should also be (at least) partially offset by a reduction in the wind-forced component of flow.

Associated with the increased outflow through Cabot Strait, and supported by the observations of a seasonal cycle in along-shelf transport on the Scotian Shelf (Drinkwater _et al._, 1979) and in the Gulf of Maine (Smith, 1983; Brooks, 1985), there should also be increased transport towards the southwest along the Scotian Shelf and in a counterclockwise sense around the Gulf of Maine. Combined with the influences of a stronger Labrador Current, the result is a general increase in the dominant component of residual circulation (with the exception of tidally-driven circulation in the Gulf of Maine region) in east coast shelf waters.

Associated with earlier seasonal snow-melt, the spring/early-summer maximum in run-off and the resulting salinity minimum and transport maximum along the coast should occur earlier in a warmer climate. Available models (Manabe and Stouffer, 1980) suggest that an advance of up to several weeks in the timing of the seasonal freshwater pulse in eastern Canadian waters is possible under doubled atmospheric CO_2 concentrations.

(ii) Stratification

As discussed in Section 3 for the open ocean north of about 45°N, there should generally be warmer and thinner mixed layers and increased stratification in the seasonal thermocline in the coastal and shelf waters off eastern Canada. The main factors contributing to this general change are a reduced rate of wind energy input across the sea surface and increased buoyancy input associated with the increases in along-shelf advection, precipitation minus evaporation (north of about 45°N), coastal run-off, net solar radiation and air temperature. Since the effects of reduced wind mixing and increased surface buoyancy input reinforce each other, the changes in stratification may be particularly significant, although their magnitude should have substantial regional and seasonal variations.

The primary regional variation in the change in summertime stratification may be a smaller increase over the Scotian Shelf and Gulf of Maine than elsewhere, associated with reduced precipitation minus evaporation over most of these areas (i.e., south of about 45°N) and a smaller increase in the freshwater discharge into the Gulfs of St. Lawrence and Maine than into more northern waters. Furthermore, over the Labrador Shelf and the

Grand Banks the stratification will also depend strongly on changes in the density of the Labrador Current and in the buoyancy input associated with sea ice and icebergs entering the region from the north and subsequently melting. Since the effects of increased northern precipitation and run-off and higher water temperatures on the surface buoyancy input may be opposite to those of a reduced input of sea ice, the net effect of these additional influences is unclear.

Wintertime (convective) mixed layers are also expected to be warmer and shallower in a milder climate scenario, except in areas with reduced wintertime ice cover (such as the Gulf of St. Lawrence and Labrador Shelf) where the exposed ocean may give up more heat to the overlying atmosphere resulting in deeper mixed layers. These changes may be pronounced since the climate-change scenarios predict maximum atmospheric changes in winter. Associated with reduced convection and shallower mixed layers in winter, there should generally be reduced water mass modification, including reduced formation of intermediate layer water (e.g., Maine Intermediate Water in the Gulf of Maine). However, in regions such as the Gulf of St. Lawrence and Labrador Shelf which are usually ice-covered in winter, any reductions in the extent or duration of the ice cover may allow increased water mass modification and increased formation of intermediate layer water.

It is also worth noting the relation of the expected changes in run-off (and hence stratification) to changes which have occurred in the seasonal cycle of freshwater discharge into the Gulf of St. Lawrence (Bugden _et al._, 1981) and Hudson Bay (Prinsenberg, 1980, 1983) due to hydroelectric power development. The anticipated increased run-off associated with doubled CO_2 concentrations in the atmosphere is in the opposite sense to the reduced springtime and summertime run-off associated with freshwater regulation, but in the same sense as the increased falltime and wintertime run-off.

(iii) Cross-shelf exchange

The net changes expected in cross-shelf exchange are uncertain and vary with region, since the relative importance of the various contributing processes has regional variations and the likely changes in these processes are often in opposite senses. The stronger Labrador Current should result in increased exchange across the shelf-break of the Labrador Shelf, the Grand Banks, and Flemish Cap associated with increased current instability and eddy formation. Increased exchange should also follow from a tendency for stronger cross-shelf residual circulation with increased along-shelf flow, partially offset by the tendency towards reduced frictionally-induced circulation with increased stratification. Similarly, an enhanced buoyancy-driven circulation in the Gulf of St. Lawrence, on the inner Scotian Shelf and in the Gulf of Maine should lead to increased horizontal exchange associated with more eddy variability. On the other hand, the weaker and less variable Gulf Stream should tend to reduce Gulf Stream eddy-induced exchange across all of the shelf regions except for the southern Labrador Shelf; there, a northward excursion of the North Atlantic Current into the southern Labrador Sea could lead to increased eddy influences on the Shelf in spite of fewer and less energetic eddies. A northward shift in the Gulf Stream position between 50° and 70°W, which cannot be ruled out, could lead to similar increased eddy and Slope Water influences on the southern Grand Banks, the Scotian Shelf and the Gulf of Maine.

Associated with reduced wind stress levels, there should be a general tendency towards reduced wind-induced exchange between shelf and offshore waters throughout the region due to reduced Ekman transport and variability, reduced shelf-break upwelling and reduced wind-induced Slope Water intrusions into the Gulf of Maine. In some cases, however, such as on the Labrador Shelf in spring, reduced Ekman transport may instead allow increased offshore (buoyancy-driven) surface flow and hence increased exchange.

Prediction of the net changes in cross-shelf exchange requires a detailed quantitative examination of the factors involved. The preliminary suggestion is that the greatest changes might be increased cross-shelf exchange on the southern Labrador Shelf and decreased exchange on the outer portions of the Scotian Shelf and Gulf of Maine.

(iv) Frontal positions

The fronts in the coastal and shelf waters off eastern Canada can generally be classified as coastal (associated with buoyancy-driven coastal currents such as on the inner Labrador Shelf, along the Gaspé Peninsula and on the inner Scotian Shelf), tidal (maintained primarily by tidal mixing such as in the Bay of Fundy, off southwestern Nova Scotia and over Georges Bank in summer) and shelf-break (such as along the edges of the Labrador and Scotian Shelves, around the Grand Banks and along the shelf edge off the Gulf of Maine). Roughly speaking, the tidal and shelf-break fronts occur over abrupt changes in depth and hence one might expect that their positions will not be changed substantially by the (relatively small) changes in driving force expected with doubled atmospheric CO_2, while the coastal fronts appear to be less controlled by topography and hence may undergo greater displacements.

The tendencies should be that the coastal fronts will be displaced farther offshore with increased coastal run-off, and the shelf-break fronts will be displaced slightly farther off the shelf with a stronger Labrador Current and increased outflow from the Gulf of St. Lawrence. Since the position of the tidal fronts is generally considered to result from a balance between tidal mixing (assumed unchanged) and solar insolation, small shifts in their position towards shallower water and a decrease in the areal extent of vertically well-mixed areas can be expected. The changes in two other factors influencing tidal front positions - reduced wind mixing and increased buoyancy input from run-off and precipitation - reinforce each other as discussed earlier, and could amplify the tendency towards a reduction in the summertime extent of vertically well-mixed areas such as found in the Gulf of Maine region.

There may be more significant changes in frontal positions that are strongly affected by ice cover. The Labrador Shelf and the Gulf of St. Lawrence are the most probable area for such changes, but there is little information on the relation of the fronts there to ice cover.

(v) Water mass properties

The water mass properties of east coast shelf waters depend on the heat and hydrological cycles in distant regions and on a global scale, as well as on the local advection, exchange, and stratification discussed above. Associated with the global increase in air temperature and the intensified global hydrological cycle, there should be a general warming and freshening of the shelf waters off eastern Canada. The magnitudes of the warming and freshening, however, should vary regionally depending on the changes in the dominant local physical oceanographic processes. The increased transport of the relatively cold Labrador Current may result in a smaller temperature increase over much of the eastern Canadian shelves than elsewhere at comparable latitudes, but this may be offset by a larger temperature increase of the source waters in more northern latitudes. On the other hand, the reduced salinity and increased transport of the Labrador Current and the increased discharge into the Gulf of St. Lawrence should result in an enhanced reduction in salinity over most of the region, particularly on the Labrador Shelf and the Grand Banks and in the Gulf of St. Lawrence. Thus, it appears that the changes in the physical oceanographic regime off eastern Canada will be such as to amplify the tendency in mid to northern latitudes toward lower ocean salinities, but it is not clear whether they will partially offset or amplify the global tendency toward higher ocean temperatures.

(vi) Sea ice

In a warmer climate associated with doubled atmospheric CO_2 concentrations, there should be a general tendency towards more year-round ice-free waters, accompanied by later freeze-up in fall and earlier break-up in spring and summer in those areas which continue to freeze over during winter. In the model prediction of the changes in seasonal sea ice for a 4 x CO_2 scenario shown in Fig. 4.1, all sea ice disappears from the Northern Hemisphere in late summer. With doubled CO_2 concentrations and a more realistic treatment of the moderating effects of oceanic heat transport, the changes in sea-ice cover should be less drastic. Although these more realistic simulations have not been done to date, we can appeal to present-day interannual variability for intuition into possible changes in sea-ice cover in the areas of interest. The ice edge limits shown in Figure 6.3 for the first week in April are representative of the seasonal maximum ice cover. There are large interannual changes as evidenced by the range in ice extent. One extreme shows an ice-free Gulf of St. Lawrence, while the other shows the Grand Bank partly covered with sea ice. Since the predicted changes in climate under a 2 x CO_2 scenario are of the order of natural interannual variations, it is likely that there will be a significant shift in mean conditions towards the minimum ice-cover conditions observed in recent times.

Regional changes in surface salinity, in wind stress, in the strength of ocean currents and in the seasonal storage of heat in the upper layers of the ocean should also influence the formation of sea ice. In the Gulf of St. Lawrence in particular, there should be the competing effects of reduced surface salinity favouring increased freezing, and warmer air temperatures and reduced wind mixing (and hence heat extraction) in fall favouring reduced freezing. For the Gulf, a mixed layer model (G.L. Bugden, pers. comm., 1985) suggests that, under altered atmospheric conditions consistent with the changes in Table 1, freeze-up will be later by a few weeks, the depth of convection at freeze-up (and hence the amount of water mass formation) will be reduced, and there will be an increased occurrence of winters without ice cover. The results of this detailed computation agree with the intuitive expectations discussed above.

7. CONCLUDING REMARKS

Climatic changes due to increased atmospheric CO_2 have not yet been conclusively observed. The predicted global temperature increase (0.3-0.4°C since 1900) could be obscured by natural climate fluctuations

(Broecker, 1975; Hansen et al., 1981). However, estimates of the natural fluctuations based on oxygen isotope analysis of a Greenland ice core (Broecker, 1975) and attempts to include the influences of other man-made trace gases and of volcanic and solar variability in modelling efforts (Ramanathan et al., 1985; Hansen et al., 1981) both indicate that the predicted CO_2-induced warming should rise above the level of expected natural fluctuations before the end of this century. Although present predictions of the effect of increased CO_2 on climate are based on models which have some obvious weaknesses, the possibility of significant global warming has been clearly demonstrated.

The major changes in atmospheric conditions predicted by present general circulation models for a doubling of atmospheric CO_2 concentrations are: a mean global surface warming of order 3°C with the largest changes in winter at high latitudes; an intensified hydrological cycle with increased net freshwater input to the ocean at equatorial and high latitudes but decreased net freshwater input in subtropical latitudes; and reduced surface wind stress.

Some of the changes which may occur in global oceanic conditions over the next century are: reduced sea-ice extent and higher sea level; reduced transport and variability (including eddy activity) in large-scale wind-driven current systems such as the subtropical and subpolar gyres; possible poleward movement of large-scale subpolar fronts associated with relaxation of the subtropical gyres; increased transport and eddy variability in buoyancy-driven coastal currents in middle and northern latitudes; a warmer and thinner surface mixed layer over most of the ocean; a fresher mixed layer at latitudes poleward of 45°N but somewhat increased surface salinities in subtropical latitudes; and reduced coastal, equatorial and shelf-break upwelling. Changes in high-latitude deep convection are both important and difficult to predict with any certainty. We speculate that the initial (decadal) response to altered atmospheric conditions will be a reduction in deep water formation rates, but we also note that this topic is particularly in need of further research.

In the waters off eastern Canada, we expect that the primary changes will be: warmer ocean temperatures; reduced surface salinities; increased along-shelf residual currents associated with a stronger Labrador Current and increased freshwater discharge into the Gulf of St. Lawrence; increased stratification and thinner surface mixed layers, except in regions of reduced ice cover where mixed layers may deepen in winter and spring; reduced areal extent of ice cover on the Labrador Shelf and in the Gulf of St. Lawrence; earlier seasonal snow-melt and ice break-up; later seasonal freeze-up; reduced Gulf Stream eddy activity; and a more northward excursion of the North Atlantic Current into the southern Labrador Sea.

At this time it is not possible to give reliable quantitative predictions. An obvious long-range approach to remedying this situation is through the development of more realistic coupled atmosphere-ocean-cryosphere general circulation models. Some improvement of these models can probably be accomplished by going to larger, faster computers that allow finer resolution and more complete physics to be incorporated. However, the parameterizations necessary in such models, and the data sets necessary to test them, require substantial development before reliable results can be obtained, meaning that process-oriented studies and observational programs must also be carried out.

Some of the inadequacies in present numerical models of atmospheric circulation (pointed out by the numerical modellers themselves) are:

(i) air-sea interactions are poorly parameterized;

(ii) the ocean models which are coupled to these models are generally highly idealized;

(iii) improved understanding of what determines humidity and cloud amount, type, and height is required (a 3% increase in low-level clouds could offset the influence on mean temperature of doubling the CO_2 concentration);

(iv) better parameterizations of the influence of humidity and clouds on both short-and long-wave radiation are required;

(v) studies of the albedo, formation and retreat of sea ice are required; and

(vi) in general, models must be developed which reproduce not only the net meridional heat flux, but also give the correct partitioning between the various components (i.e., predictive capability requires the right answers for the right reasons).

Oceanic numerical models suffer from problems analogous to those of their atmospheric counterparts. Present models are not able to reproduce accurately the thickness of the main thermocline and we do not yet fully understand what determines the position, strength, and variability of such major currents as the Gulf Stream, the North Atlantic Current and the Labrador Current (the Labrador Current does not even appear in most oceanic general circulation models). Clearly, before we can expect to produce reliable predictions of changes in the ocean we must at least be able to model the present state. To accomplish this goal, the parameterizations of subgrid-scale mixing, deep convection processes, eddies, air-sea interaction and bottom stresses must be improved.

If we are to achieve the goal of providing reliable climate predictions, it is clear that the interplay between observations and modelling must play a critical role. This is true for all three components of the atmosphere-ocean-cryosphere system. Observations do not only present a picture of the present state; they often reveal unforeseen aspects of the processes determining this picture which are critical to the ongoing development of our understanding and, of course, they must supply the final check on the validity of model predictions. Monitoring of the atmosphere-ocean-cryosphere system must be maintained over the next several decades in particular in order to observe any predicted changes, if and when they occur, and <u>all</u> data must be archived with "history files" containing information on quality control and processing to date. For the ocean, some appropriate indicators of large-scale changes are sea level, sea-ice extent (including annual and interannual variations), the large-scale thermohaline properties (such as might be determined by ship-of-opportunity programs), and the transports and positions of major ocean currents.

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APPENDIX: A BUCKET MODEL OF THE ATMOSPHERE-OCEAN INTERACTION PROBLEM IN CLIMATE MODELS (A THOUGHT EXPERIMENT ON TIME SCALES)

To aid in our understanding of the role of the ocean in the climate system, it is interesting to extend the leaky-bucket analogy (Section 2) of Hansen et al. (1981). Although it must be kept in mind that this analogy is limited, we can derive further insight from it and emphasize some of the uncertainties with which we are faced.

Figure A.1 illustrates an (oversimplified) analogue of the flow of heat through the ocean-atmosphere system. It is composed of only two buckets containing water (the working fluid - analogous to heat in the ocean-atmosphere system) connected by a valve. The container on the left (C1) represents the atmosphere. The tap T1 represents the source of incoming short-wave solar radiation, and T2 represents the loss by long-wave radiation. C1 is connected to a much larger container (C2) representing the ocean (the container is much larger to reflect the large heat capacity of this component) through a valve (V1) (the ocean-atmosphere interface) which controls the time scale of adjustment between the two regions.

This simple configuration is analogous to atmospheric models which include only a mixed layer ocean. Increasing the CO_2 content of the atmosphere is analogous to turning down T2 (reduction of the Earth's mean albedo would correspond to turning up T1 but we will not discuss feedback mechanisms here). The main point to be made from this analogy is that the adjustment to a change in atmospheric CO_2 content depends on the exchange rate represented by V1.

If V1 were closed then turning down T2 (analogous to the effect of increased atmospheric CO_2) would result in an increase in water level in C1

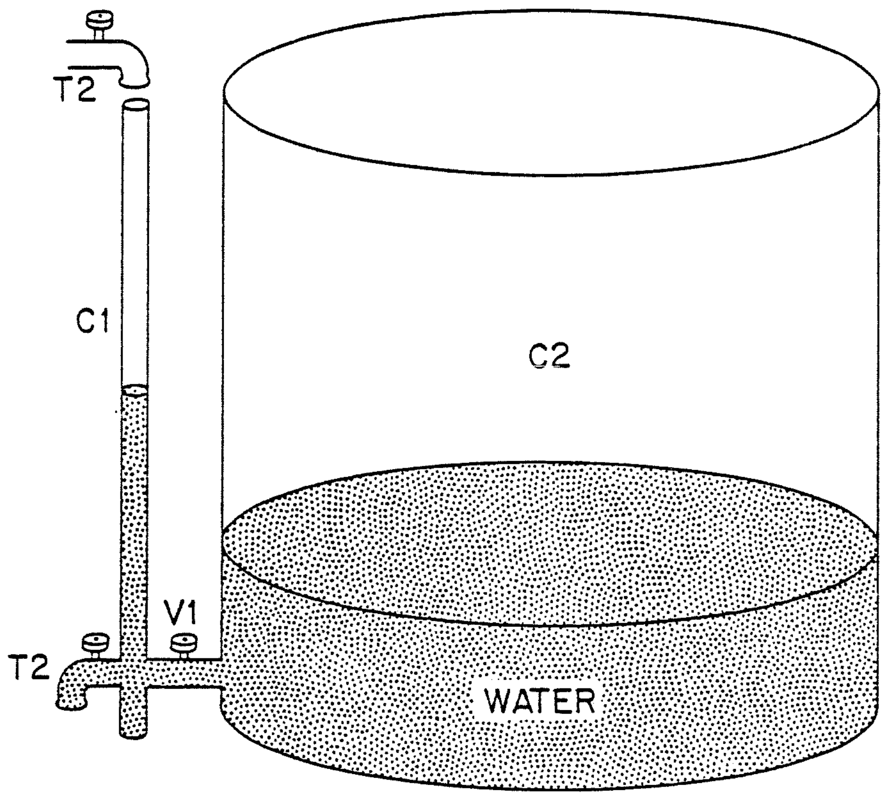

Fig. A-1. Illustration of the components of the bucket analogy. Note that heat loss to other latitudes is neglected consistent with mixed layer ocean models - a serious limitation of the analogy and the numerical models.

at a rate determined by the surface area of C1 and the excess of flow through T1 over that through T2. The level would rise until the pressure at T2 was sufficient to drive flow out at the same rate as it is input through T1. Due to the small surface area of C1 the water level would rise rapidly to its new equilibrium position.

If V1 were opened only slightly, the water level in C1 would adjust to turning down T2 essentially as before, and the level in C2 would subsequently slowly rise to that in C1 due to the pressure difference across V1. Conditions in C1 would thus be affected little by exchange with C2.

On the other hand if V1 were opened enough to allow equilibration of the levels in C1 and C2 over the time scale on which the level in C1 adjusts when V1 is closed, the time scale of adjustment in C1 would be greatly increased. In this case the rate of adjustment of the water level in C1 would be dictated by the total surface area of C1 and C2 together (i.e. by the combined heat capacity of the model ocean and atmosphere). Noting that the heat capacity of the entire atmosphere is about the same as that of the upper 3 m of the ocean at 10°C, and that at least the upper several hundred to a thousand meters of the ocean is renewed on no longer than a decadal time scale, it is clear that the ocean's influence can result in a significant time delay of the atmosphere's response to a sudden change in CO_2 content (numerical model results suggest that this delay is of order 10-25 years). However, just as the equilibrium level in C1 is nearly independent of the exchange through V1, the equilibrium temperature of the atmosphere (averaged over a year and over the globe) is not strongly dependent on the amount of heat stored in the ocean. The major role of the ocean in determining the mean atmospheric temperature is as a source of moisture (one of the feedbacks neglected in the above discussion). This is one reason why swamp models of the ocean (a representation of the ocean as a source of moisture with zero heat capacity) result in reasonable estimates of the equilibrium temperature of the atmosphere. However, they probably will not give good estimates of the adjustment process or of the changes in atmospheric weather patterns.

The simple analogy discussed above emphasizes the need to know exchange rates between the atmosphere and the ocean. While chemical tracers and modelling efforts have provided useful information on this subject, much more work is needed to be able to predict accurately changes in ocean circulation which will be associated with changing CO_2 concentrations.