Three-Dimensional Observations of a Deep Convective Chimney in the Greenland Sea during Winter 1988/89

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(Abstract received 1 March 1995, in final form 20 February 1996)

ABSTRACT

All available temperature data, including moored thermistor, hydrographic, and tomographic measurements, have been combined using least-squares inverse methods to study the evolution of the three-dimensional temperature field in the Greenland Sea during winter 1988/89. The data are adequate to resolve features with spatial scales of about 40 km and larger. A chimney structure reaching depths in excess of 1000 m is observed to the southwest of the gyre center during March 1989. The chimney has a spatial scale of about 50 km, near the limit of the spatial resolution of the data, and a timescale of about 10 days. The chimney structure breaks up and disappears in only 3–5 days. A one-dimensional vertical heat balance adequately describes changes in total heat content in the chimney region from autumn 1988 until the time of chimney breakup, when horizontal advection becomes important. A simple one-dimensional mixed layer model is surprisingly successful in reproducing autumn to winter bulk temperature and salinity changes, as well as the observed evolution of the mixed layer to depths in excess of 1000 m. Uncertainties in surface freshwater fluxes make it difficult to determine whether net evaporation minus precipitation, or ice advection, is responsible for the observed depth-averaged salinity increase from autumn to winter in the chimney region. Rough estimates of the potential energy balance in the mixed layer suggest that potential energy changes are reasonably consistent with turbulent kinetic energy (TKE) production terms. Initially the TKE term parameterizing wind forcing and shear production is important, but as the mixed layer deepens the surface buoyancy production term dominates. The estimated average annual deep-water production rate in the Greenland Sea for 1988/89 is about 0.1 Sverdrups, comparable to production rates during the 1980s and early 1990s derived from tracer measurements. The location of the deep convection observed appears to be sensitively linked to the amount of Arctic Intermediate Water (AIW) present from autumn through spring. Although AIW is an important source of salt for the surface waters, too much AIW overstratifies the water column, preventing deep convection from occurring.

1. Introduction

Deep ocean convection occurs in only a few places in the world and is sensitive to changes in the thermohaline circulation and surface forcing at those locations. The rate at which it occurs largely determines the residence time of the deep ocean, and therefore a better understanding of the process is important to our understanding of global climate change.

The deep convective process in its entirety occupies a wide range of spatial scales (Gascard 1991). Preconditioning may be linked to mesoscale eddies (Johannesen et al. 1991) and to the gyre-scale circulation (Killworth 1979), as well as to bathymetric features (Hogg 1973). During deep convection, “plumes” are the smallest [O(1 km)], most intense convective signals that have been observed (Schott and Leaman 1991; Schott et al. 1993) and appear in numerical models (Brugge et al. 1991; Jones and Marshall 1993; Paluszewicz et al. 1994). A larger [O(50 km)] “chimney” scale is believed to encompass the plumes; features around this size have been observed at several deep convective sites (MEDOC Group 1970; Gascard 1978; Gordon 1978; Killworth 1979; Gascard and Clarke 1983; Johannesen et al. 1991). Observations of the adjustment of convected waters suggest breakup into mesoscale eddies through baroclinic instability (Killworth 1976; Gascard 1978; Gascard and Clarke 1983).

The occurrence of deep ocean convection in the Greenland Sea has long been predicted (Nansen 1906) but rarely observed. Evidence from tracers suggests that ventilation rates in this region have decreased and since 1983 average about 10% of the rates in the 1970s.
(Schloesser et al. 1991; Rhein 1991, 1994). The lack of observations is in part due to the harsh environmental conditions under which convection occurs, in part due to the short spatial \(O(\text{km})\) and temporal \(O(\text{days})\) scales over which recent evidence suggests the strongest signals occur and in part due to the recent reduction in annual deep-water production rates.

From summer 1988 to summer 1989, an extensive monitoring program occurred as part of the international Greenland Sea Project. This program included seasonal hydrographic surveys and the first moored temperature, salinity, velocity, and tomographic time series in the region. Despite relatively heavy coverage, observations of deep convection by the thermistor and hydrographic data were sparse, only capturing plume-scale events. Sixteen deep CTD casts were made within the gyre close to the time of maximum convection, but only one showed a mixed layer extending to 1500-m depth. From the moored instruments, a single plume reaching at least 1400-m depth was observed at the center of the gyre for one day by upward looking acoustic doppler current profilers (ADCP). This led investigators to conclude that homogeneity of the water column to depths greater than 250 m was sporadic and small scale (Schott et al. 1993). Estimating the evolution of the large-scale temperature field using only the point measurements\(^1\) is difficult due to the short internal deformation radius at high latitudes.

By spatial averaging across many correlation scales, the tomographic data are well suited for measuring the large-scale temperature (sound speed) fields but lack the spatial resolution of point measurements. Acoustic inversions by Worcester et al. (1993) and Pawlowicz et al. (1995) present a somewhat different picture of deep convective activity in the gyre. Worcester et al. (1993) show the first year long time series with evidence of large-scale cooling penetrating to \(\sim 1000\) m on average over the 100 km distance between moorings. Pawlowicz et al. (1995) extend these results, breaking up the temperature evolution into the different stages believed to be associated with the convective process (MEDOC group 1970). They particularly concentrated on the daily temperature evolution of the mooring 4 to mooring 6 path (Fig. 1). These papers incorporated only the rays that could be tracked for the entire year, giving suboptimum but consistent vertical resolution. Acoustic inversions by Sutton et al. (1994) focused on the near-surface temperature evolution of the average ocean between mooring 1 and mooring 6 at 4-hour intervals. In doing so, acoustic normal modes were incorporated in the inversions, greatly enhancing the vertical resolution near the surface as well as reducing the uncertainty over the whole water column.

This paper presents results from inversions that for the first time estimate the three-dimensional evolution of the temperature field in the Greenland Sea at chimney scales and larger. In previous work, acoustic inversions were done separately for each mooring pair and therefore gave range-averaged temperature profiles over regions dictated by the array geometry, rather than by the convective pattern. The three-dimensional inverse allows us to examine any volume of interest within the array and, in particular, to focus on the evolution of a convective chimney about 30 km to the southwest of the central mooring where cooling penetrates deepest. These solutions indicate that the sparse deep convective activity observed by the point measurements is more due to where and when they sampled rather than the lack of significant deep convective activity in the gyre.

The inversions presented here are also unique in that they combine the full acoustic dataset, including acoustic normal modes, all moored thermistors in the gyre, and hydrographic data where appropriate. Technical details of the inversion are described in Morawitz et al. (1996). Because of the complementary characteristics of the acoustics and the traditional point measurements, a proper combination of the different data types provides a powerful analysis tool. Additional advantages in using formal inverse methods are twofold. First, the different data types can be objectively combined to give optimum estimates of the temperature field. Output solutions are then consistent with all observations included to within assigned data errors. Second, the uncertainty of the estimated temperature field can be calculated.

2. Measurements

A rich set of disparate measurements were made during 1988–1989 by various investigators participating in the Greenland Sea Project (GSP) (Table 1). Twenty-seven moored thermistors were located within 110 km of the central mooring (mooring 6) (Fig. 1). Approximately half of these instruments were located on tomographic moorings 1, 3, 4, 5, and 6. Time series from thermistors and conductivity sensors on the central mooring have been reported by Roach et al. (1993). An additional mooring deployed by the University of Kiel (mooring 319) carried a ten-element thermistor string in the upper 250 m and three deeper thermistors (Schott et al. 1993). Temperature series at 90 m are shown in Fig. 2. Moored thermistors were manufactured by SeaCat (Aanderaa) with typical errors around 0.02°C (0.1°C).

Seasonal hydrographic measurements, with temperatures and salinities calibrated by the Scripps Oceanographic Data Facility, were made at the approximate locations shown in Fig. 1. Temperature profiles from the winter–spring time period will be shown in a later section. Profiles from other casts during 1988–1989

\(^1\) Point measurements refer to the hydrographic and thermistor data, which measure temperature at points in space, as opposed to acoustic data, which integrate over space.
Fig. 1. Top: Map of the Greenland Sea showing surface currents, bathymetry, Polar Front, and 16 March 1989 20% ice edge. Reciprocal transmissions occurred on the acoustic paths shown by the heavy lines; thin lines are paths with one-way travel times only, due to mooring 3 receiver failure. Mooring 2 failed completely after one month, so the dashed paths could not be used. Mooring 319 carried temperature, conductivity, and velocity instrumentation only (no acoustics). Bottom: An expanded view of the array, with locations of all the disparate data used in the inversions. Small circles denote a Seasoar track through the region and the ∇ denote deep CTD stations, both part of the SIZEX cruise in March 1989. Open circles are CTD stations occupied seasonally throughout the year.
TABLE 1. Data incorporated in inversions.

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<th>Mooring location</th>
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<th>Source</th>
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<td>K. Aagaard</td>
</tr>
<tr>
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<td>60–240, every 20 m</td>
<td>F. Schott</td>
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<tr>
<td>Valdivia 78</td>
<td>Feb–Mar 1989</td>
<td>J. Swift, J. Meincke</td>
</tr>
<tr>
<td>SIZEX (CTD, Seaosoar)</td>
<td>Mar 1989</td>
<td>O. Johannessen</td>
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<tr>
<td>ArkVI</td>
<td>May 1989</td>
<td>J. Swift, G. Budeus</td>
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Acoustic data

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<td>Other</td>
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can be found in GSP Group (1990), Bader et al. (1991), van Aken et al. (1991), Rhein (1991), and Budéus et al. (1993). Hydrographic measurements were also taken in the region in mid-March 1989 as part of the SIZEX experiment (SIZEX Group 1989). This experiment included 11 hydrographic stations and a towed Seaosoar survey, giving temperature, salinity, and pressure over the upper 250 m at the locations shown in Fig. 1. The GSP hydrographic measurements have errors of 0.002 in salinity and 0.002°C in temperature, while the SIZEX cruise has larger errors of 0.04 and 0.005°C.

Examples of acoustic receptions have been presented in Worcester et al. (1993), Sutton et al. (1993), Jin et al. (1994), and Morawitz et al. (1996). We refer the interested reader to those articles.

Additional datasets used in the analysis here are ice cover derived from satellite microwave radiometer (SSM/I) data and heat flux and wind stress data provided by the British Meterological Office (BMO) data-assimilation model (Bell and Dickinson 1987). SSM/I grid points are spaced 25 km apart, but this resolution can be degraded due to an intermittent 25 km earth location error in the SSM/I during 1988–1989 (Roach et al. 1993). The NASA algorithm ice concentration product has an accuracy of approximately 5% (Cavaliere 1992) in areas of high ice concentration. We are concerned here with thin, newly formed ice, and so this accuracy estimate is probably overly optimistic. Evidence suggests that SSM/I is sensitive to the thickness of this ice type in addition to its concentration (Grenfell et al. 1992), so that low ice concentrations could actually be higher concentrations of thin ice.

The BMO heat flux and wind stress are hindcasts based on bulk parameterizations of surface and atmospheric variables that are assimilated into the model (Bell and Dickinson 1987). Estimates are made every 3 hours at grid points spaced 0.9375° in latitude and 0.75° in longitude (approximately 105-km spacing meridionally and 22 km zonally in the Greenland Sea). A correction scheme similar to that proposed by Pawlowitz et al. (1995) has been applied to account for incorrect ice limits and sea surface temperatures during autumn. The long-term (monthly) biases in the corrected heat fluxes are estimated to be ~50 W m⁻² and are probably largest in November and December, the time of maximum ice cover. Because of the horizontal smoothing, the model will not be able to resolve the shorter horizontal scale surface flux variability that can occur at boundaries such as the ice edge. BMO wind speeds differed from shipboard observations by 1 ± 3 m s⁻¹, with small directional differences over the range 0–16 m s⁻¹.

3. Chimney timescales and space scales

The Greenland Sea is a topographically confined basin (Fig. 1) bounded by the East Greenland slope to the west and Jan Mayen Fracture Zone and Mohns Rise to the south and east. Because the water column is weakly stratified, the bathymetry is important to the dynamics and tends to define the Greenland Sea gyre, as is evident in contours of dynamic topography (Clarke et al. 1990), in current meter data (K. Aagaard 1994, personal communication), and in numerical models (Legutke 1991). The East Greenland current to the west has a cold, fresh, surface layer of polar water originating north of Fram Strait, overlying a warm and saline subsurface temperature maximum near 200 m. This temperature maximum is part of the Arctic Intermediate Water (Swift 1986) and has its origin in the North Atlantic. The Jan Mayen Current breaks from the East Greenland Current and heads eastward following the Jan Mayen Fracture Zone, and the gyre is completed by waters continuing northward following the western bank of the Mohns Rise. Within the Greenland
Basin beneath the 50-m surface layer, intermediate waters near the gyre center are colder, denser, and less stratified than waters at the same depth on the periphery. Isopycnals therefore slope downward to the warmer and saltier perimeter, consistent with the net cyclonic circulation (Quadfasel and Meincke 1987).

In this context, the center of the array domain (Fig. 1) will be referred to as the center of the Greenland Sea gyre. The northern, southern, and western perimeters of the functional array, close to the edge of the abyssal plain, are generally in the vicinity of the gyre perimeter as indicated by the 1988–1989 temperature data (Worchester et al. 1991; Pawlowicz et al. 1995). For the purposes of this paper, those regions of the array will be equated with the gyre perimeter.

In the summer, the near-surface layer is warmed by the sun and is fresh due to the previous winter’s ice melt and/or advection of polar waters across the Polar Front. During autumn, surface waters cool rapidly as air temperatures drop and solar insolation decreases, but the water column remains stable due to the low salinity of the surface layer. Cooling is confined to a thin layer, which rapidly approaches the freezing point. In late November and December 1988, satellite im-
agery shows the ice edge moving in from the west so that by January the majority of the Greenland Sea is ice covered. Salinity time series show a steady increase in the upper ocean as surface waters become denser through a combination of brine rejection associated with ice formation and entrainment of Arctic Intermediate Water from below (Roach et al. 1993). This results in a gradual erosion of the pycnocline. In February and March 1989, the classic “Is Odden”--”Nordbukta” ice formation, first recognized by Norwegian whalers, occurs. A large ice-free area forms (Nordbukta) surrounded by ice (Is Odden). It is in this area and over this time period that a convective chimney is observed in the three-dimensional temperature fields from inversions described in detail in Morawitz et al. (1996) and briefly summarized in the appendix.

The convectively formed waters that will be the focus of our discussions are believed to replenish deep waters in both the Arctic and North Atlantic basins. The less dense waters are part of the Arctic Intermediate Water (AIW) found in the upper few hundred meters in the gyre. The AIW in the Greenland Sea combines with intermediate waters from the Arctic Ocean outflow and from the Iceland Sea and overflows the Denmark Strait (Peterson and Rooth 1976; Swift et al. 1980; Aagaard et al. 1985, 1991), although this has been recently questioned by Mauritzen (1993). Some of the underlying Greenland Sea Deep Water may reach the North Atlantic through the deeper Faroe–Shetland Channel (Mauritzen 1993) after mixing with deep waters originating in the Eurasian Basin and then advecting into the Norwegian Basin through passages in the Jan Mayen Fracture Zone (Aagaard et al. 1985). Greenland Sea Deep Water below roughly 1000–1500 m is confined to the north of the Greenland–Scotland ridge system, some of which renews Eurasian Basin Deep Water (Aagaard et al. 1985).

a. Plan view

Figure 3a shows the depth-averaged potential temperature between 500 and 1000 m over the array domain in mid-March 1989, using all temperature measurements except the deep hydrographic stations. Each panel is an independent inversion for the 3-day average temperature field. Before 19 March, the temperature structure is typical of the Greenland gyre, with colder, central waters surrounded by warmer water to the north, west, and south against the East Greenland slope and over Mohns Rise at 74.2°N. From historical hydrography, waters within the East Greenland current are on average 0.45°C warmer than waters in the central gyre over this depth range, so that the warm region is influenced by, but not entirely composed of, East Greenland and Jan Mayen Current waters.

Within the central gyre, waters are about 0.2°C colder than around the perimeter. On 19 March, a confined region to the southwest of the central mooring abruptly cools to less than −1.32°C. This cold core persists into the next 3-day inversion and then loses intensity over the next 6 days. As the Greenland Sea waters are the coldest waters in the region at this depth (Clarke et al. 1990), the cooling must come from above. Because of the abrupt timescale (−10 days), the cold core is suggestive of convection physics. The 50-km spatial scale is similar to what has been called the chimney scale in other regions, and we shall refer to this cold core as the chimney region and predominantly focus on this area in further discussion.

In the preceding analysis, deep hydrographic casts during the convective time period were excluded because the ocean parameterization used in the inversions is not meant to include plume-scale variability (we are only concerned with the chimney-scale and larger field here). Whereas the 3-day averaged time series and acoustics anti-alias short time and space scale plume events by temporal and/or spatial averaging, the CTD measurements sample at a single point in both time and space; if a cast captures the intense signal of a single plume, it can have a large and unrealistic effect on the solution map. This is illustrated in an alternate 10 March panel in Fig. 3b. The cold feature, delineated by the −1.32°C isotherm, is entirely due to the southernmost CTD cast on that day. This “plume cast” is the only cast taken during 1988–1989 that shows a deep mixed layer penetrating to about 1500 m (Fig. 4). From the surrounding CTDs, and the time series based on all other data, this cast appears to sample an early isolated plume before enough are present to give a measurable signal on the chimney scale. Other differences between the inversions, including and omitting the deep hydrographic casts, are within solution uncertainty. Rather than bias our solutions by only omitting this extreme cast, the inverse solutions will not include any of the CTDs taken during the convective time period.

Even though the central gyre is the most nearly homogeneous region, substantial variability still exists over short spatial scales during winter, in addition to the signal associated with the convective plume (Fig. 4). The two outlying casts, with the coldest and shallowest mixed layers, have temperature profiles similar to inverse solutions around the gyre perimeter and probably show water that has been advected into the central gyre, as additional evidence in a later section will suggest. Of the two CTD casts within the chimney region, only the plume cast shows a significantly deeper mixed layer depth than the other casts. Both casts within the chimney have a relatively homogeneous upper water column, suggestive of active mixing. The temperatures above 250-m depth in these two casts are

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2 \( r_1 = 32.785 \) is defined by Aagaard et al. (1985) to be the isopycnal separating intermediate and deep waters.
Fig. 3a. Depth-averaged potential temperature between 500 and 1000 m. The contour interval is 0.04°C. Each map is a 3-day average, centered on the day shown in the top-left corner of each panel. The colored boxes give the locations of CTD casts, with the color proportional to the average temperature of the cast between 500 and 1000 m. These casts were not included in the inversions. Bathymetry of the region is drawn in black in the first panel; bathymetry contours are labeled in Fig. 8. Three-day average velocities from current meters are shown as red arrows. The last panel shows the rms uncertainty of the inversions (in °C). The dashed box indicates the area that defines the chimney region.
warmer than the average chimney profile from the inversions, although the difference is comparable to the expected uncertainties [cold surface waters during a deep mixing event were also observed by Schott et al. (1993) within a single plume]. On 16 March, before the chimney develops, the large-scale temperature structure of the chimney region (the inverse solution) is typical of the majority of the CTD casts in the central gyre.

b. Mixed layer depth

The minimum depth of the −1.2°C isotherm serves as a surrogate for mixed layer depth and is shown preceding and during the deep convective event (Fig. 5). Initially, on 13 March, the central gyre has a bowl-like shape with relatively shallow mixed layer depths around the perimeter. Over the next 6 days the mixed layer preferentially deepens over an area \( O (50 \text{ km}) \) in diameter centered at 74.75°N, 3.5°W, reaching a maximum depth of about 1500 m. The original bowl-like shape of the gyre splits in two, probably as warmer water is advected in from the perimeter. This tongue of warmer water can be seen in the 22 March panel of Fig. 3a as well. The current meter data from mooring 5 suggest advection from the northwest when the chimney is most intense, although the data are insufficient to draw definite conclusions. After 19 March, the deepest mixed layer is confined to the chimney region. The temperature structure becomes more complicated on 28 March, when water is advected into the gyre and the chimney breaks up.

c. Vertical sections

Two vertical slices were chosen to show the time evolution of the temperature structure within the chimney (Fig. 6); their orientation is shown in Fig. 5. The zonal section intersects the mixed layer depth maximum, while the meridional transect is to the west of the deep maximum but bisects the chimney at shallower depths. The cooling initially penetrates to a broader extent in the east-west plane than in the north-south, reflecting the predominantly east-west orientation of the initial deep mixed layer trough in Fig. 6. The meridional slice through the volume indicates that a preferentially cooled region already exists on 7 March, however, at 75°N, 4.46°W, slightly north of the area that convects 9 days later. This is also slightly north of, although at the same time as, the deep mixed CTD cast that was observed. The edge of the warm intermediate water that splits the gyre in Fig. 5 can be seen appearing at about 2.5°W on 19 March (Fig. 6a), where it remains and strengthens through 28 March. The larger uncertainties near the edges of both maps reflect the poorer sampling there due to the geometry of the array.

The SSM/I ice cover concentration is plotted above each panel in Fig. 6. The original ice data resolution is at best 25 km; the data have been linearly interpolated for this display. Minimum ice cover generally overlies areas where surface waters are warmest, for example, in the vicinity of the deep mixing sites. The emerging picture is suggestive of one-dimensional vertical mixing. The flux of heat from waters below warms the upper ocean despite large surface heat loss to the atmosphere in a region where ice cover is a minimum. Unfortunately, the temporal resolution is insufficient to decide between two possible scenarios: (i) the ice cover decreases before the mixing event, leading to larger surface heat losses and thus convective instability, or (ii) a convective event brings warmer water to the surface leading to a subsequent melting. Ship observations in this area around 10 March found that despite very cold air temperatures, the surface waters were significantly warmer than freezing, and the small amount of ice present appeared to be melting (J. Swift 1990, personal communication). The potential temperature of the water column, despite vertical mixing in the vicinity of the chimney, never completely homogenizes.

d. Vertical profiles

Potential temperature profiles averaged over the chimney area are presented for most of the winter and spring to show the time evolution of the chimney (Fig. 7). In late January, following the period of heaviest ice cover, when brine rejection and entrainment have decreased the density difference between the mixed layer and the underlying water column (Roach et al. 1993), the mixed layer is \( \sim 200 \text{ m} \) deep and at a temperature of about \(-1.6°C\). Heat fluxes out of the ocean increase as the insulating ice cover decreases, but surface waters warm as heat is mixed up from below. The depth of the \(-1.2°C\) isotherm increases in abrupt steps that occur near 1, 15, and 28 February, separated by more
gradual deepening. Pawlowicz et al. (1995) invoke shear-driven entrainment to explain deepening of the mixed layer to 250 m. Schott et al. (1993), from their thermistor observations 16 km west of the northwest corner of the chimney region (m319), associate the first two events with the strong heat fluxes and southerly wind bursts that occur at those times. They could not resolve the extent of the deepening of the second event, other than to say that it extended past their intermediate thermistor at 350 m but failed to reach an instrument at 1300 m. From the inversions, cooling associated with this event reached about 600 m within the chimney. This is followed by a third event on 28 February associated with a peak in heat flux and an increase in ice formation. While ice concentrations are not large during these times, the hints of ice growth during convective time periods mean that ice could be playing a role in triggering convection, as hypothesized by Rudels (1990), although the SSM/I data are inadequate to determine whether ice forms before or after convection.

The chimney in its maturity lasts from 19 until 28 March when the water column begins to restratify, initially at 500 m, and later at shallower depths around 150 m (Fig. 7). Soon afterward the structure near the surface becomes more complicated as warmer waters move into the region. At about the time of chimney breakup acoustic normal modes could no longer be identified, decreasing the inverse resolution in the upper water column. As normal modes are sensitive to small perturbations in the sound speed profiles near the surface, their sudden disappearance is consistent with the near-surface sound speed structure becoming more complicated. The increased variability in the inverse solutions is not due to the loss of normal modes because the basic characteristics of the solutions remain the same whether they are used or not. Evidence from current meters on the central mooring (Fig. 14) also suggests a much more variable and energetic velocity field around this time. The chimney breakup will be discussed in more detail in section 6.
Fig. 5. The minimum depth of the $-1.2^\circ\text{C}$ isotherm over the array serves as an indicator of mixed layer depth. Contours of the mixed layer depth are drawn below to indicate the latitude-longitude position of the chimney and its orientation. The bold lines show the locations of the transects displayed in Fig. 6.
Fig. 6a. Potential temperature for consecutive 3-day averages, shown as vertical slices along 74.75°N. The length of the slice is 157 km. Contour intervals are 0.1°C. Average SSM/I ice concentrations for each 3-day period are shown overhead as percent ice cover. Estimated uncertainty is shown in the bottom-right panel (in °C).
Fig. 6b. Same as Fig. 6a but for a 157-km north–south slice at 4.46°W.
Fig. 7. Bottom: Time evolution of potential temperature profiles averaged over the chimney region. Each profile is a 3-day average, which has been linearly interpolated in time. Contour intervals are 0.1°C. Typical rms uncertainty (°C) is shown to the right. Plotted above are the SSM/I daily averaged ice cover over the area of the chimney, daily averaged total heat flux over the chimney from the BMO model, and daily averaged wind stress over the chimney from the BMO model, respectively.

e. Salinity

The chimney observed in the temperature field is also evident in salinities from Seasoar transects made across the Greenland gyre in mid-March 1989. The highest salinities over the upper 100 m occur in an $O(50 \text{ km})$ diameter patch to the south-southwest of the central mooring (Fig. 8). Seasoar data was included in the inversions as sound speed, but because sound speed is only weakly dependent on salinity, this salinity data are essentially independent of the inversions. The presence of relatively saline water near the surface during the cold winter season is a strong indication of a likely convective region. In-
creased salinity in the upper 200 m can be explained by entrainment from the underlying halocline, as was observed in salinity measurements on the central mooring (Roach et al. 1993). While Seasoar data are limited to the ship tracks, the salinity signal led the SIZEX investigators to suggest a deep convective region would be centered approximately 20 km to the south of the central mooring over a patch on the order
of 60 km in diameter (SIZEX Group 1989). This area includes the chimney site previously discussed.

4. Heat and salt budgets

The temperature field, the near-surface salinity data from the SIZEX cruise, and the ice data strongly suggest that vertical mixing is dominant within the chimney. This hypothesis is supported by looking at changes in temperature over successively deeper depth intervals and finally comparing surface heat fluxes over the chimney to changes in heat content of the entire water column (Fig. 10). A simple one-dimensional model is then constructed to compare to observed depth-integrated salinity and temperature and to the observed time evolution of the mixed layer.

The average temperature of the upper 100 m increases as the mixed layer deepens (Fig. 10). The period of largest heat gain corresponds to the time of largest heat loss in the underlying water column between 100 and 500 m. This depth range contains the largest temperature gradient associated with Arctic Intermediate Water (Fig. 7), and the largest upward flux of heat occurs as mixing penetrates through this range. Although there is evidence of warming in the layer 100–500 m as cooling penetrates into the next deepest layer, specifically after the deepening event on 15 February and through the event at the beginning of March (see Fig. 7 for reference), the signal is smaller, primarily because the vertical temperature gradient decreases below 500 m. Over progressively deeper ranges cooling occurs later, as expected for a surface-forced process. There is no evidence that cooling penetrates past 2000 m.

In a one-dimensional process, neglecting horizontal advection, diffusion, and mixing of heat, the average temperature change $\partial \Theta / \partial t$ from the surface to a depth where vertical fluxes vanish, $-D_{\alpha}$, can be related to the net surface flux of heat out of the water column, $B_{\alpha}(t)$, so that

$$\frac{B_{\alpha}(t)}{c_{p}\rho D} = \frac{1}{D} \int_{-\infty}^{0} \frac{\partial \Theta(z,t)}{\partial t} \, dz, \quad (1)$$

where $c_{p} = 3988 \text{ J kg}^{-1}\text{C}^{-1}$ is heat capacity and $\rho = 1028.07 \text{ kg m}^{-3}$ is the density of sea water. Both are taken to be constant for simplicity. Here $D$ is chosen to be 2000 m. The time integral of both terms in (1) is shown in the bottom panel of Fig. 10. From midwinter until the time the chimney breaks up around 25 March, the agreement between the two is much better than could be expected given the uncertainties in the BMO model. Pawlowicz et al. (1995) also found the range-averaged heat content on the 4 to 6 path, the path most in common with the chimney domain, to be consistent with a one-dimensional heat balance.

As previously stated, the salinity budget in the upper ocean plays an important role in ocean convection at high latitudes. There is a net increase in salinity in the upper 1000 m of the central Greenland Sea during late fall and winter 1988/89 that can be explained by the brine rejection associated with 50 cm of ice growth (GSP Group 1990). Roach et al. (1993) hypothesized that ice was advected southward out of the central gyre to account for the net salinity increase, and Pawlowicz et al. (1995) argued that advection of ice out of the convective region is crucial to determining the depth of convection. We revisit the question of whether or not advection of ice is essential to explain the depth of convection observed during 1988–1989 using a simple one-dimensional mixed layer model previously appearing in Clarke and Gascard (1983), Clarke et al. (1990), and Sutton et al. (1993).

The model is nonpenetrative and stability driven. Nonpenetrative models have been shown to describe the evolution of the mixed layer effectively in previous convective studies (Anati 1971; Killworth 1979), although they are obviously oversimplified. The model is initialized with vertical profiles of temperature and salinity. Daily averaged net and latent surface heat fluxes from the BMO determine the cooling and evaporation rates. The model is time-stepped forward and the surface fluxes applied, modifying the density at the surface. If the overlying water becomes denser than the water beneath it, the profile is vertically mixed just enough to restore stability. If the surface water reaches the freezing point, ice is formed, rejecting salt into the surface layer. As the mixed layer deepens, heat is entrained from below. If ice is present, this heat is used to melt the ice. Ice melt is spread somewhat aphysically over the entire mixed layer by invoking wind stirring, which tends to give more realistic results (Martinson 1990).

The model was run twice, using as initialization two hydrographic casts in the central gyre from autumn
Fig. 9. Daily averaged total ice concentrations from SSM/I. Red arrows give daily averaged wind stress from the BMO model.
Fig. 10. Depth-averaged potential temperature over various depth ranges. Rms uncertainties are shown by the vertical bar at the beginning of each curve. In the bottom panel (0–2000 m), the average temperature (appropriately initialized) due to surface heat fluxes from the BMO model is shown as a dashed line. The shaded region denotes the temperature error for up to a ±50 W m\(^{-2}\) error in the BMO heat fluxes that is independent month to month.

1988. The measurements were made just before the ice edge moved into the central gyre and are therefore good initialization points for temperature and salinity budgets. Latent and total heat fluxes averaged over the box shown in Fig. 11 were used to force the model through evaporation and cooling (the latter part of the time series is shown in Fig. 12). The average evaporation rate as calculated from the latent heat flux is 0.002 m/day, consistent with the net freshwater loss used by Clarke et al. (1990) in their model runs for the region.

To compare our model results to observations, we average temperature and salinity from the surface to 1800 m, which is just below the maximum depth of convective cooling, so that vertical heat and salt fluxes to greater depths are negligible (1800 m was chosen instead of the 2000-m depth used in previous heat con-
only a slight variability in salinity. This is in contrast to changes in salinity at shallower depths, where water salinity can increase by $O(0.4)$ from autumn to winter, emphasizing the importance of entrainment of salt into the mixed layer from AIV water below.

The depth-averaged temperature and salinity on 10 March from the two model runs compare well with the two coldest and freshest wintertime casts, in which cooling penetrates to depths of about 1700 m and 1200 m. The third winter cast has the shallowest mixed layer in Fig. 4. It has temperature characteristics typical of perimeter waters and is inconsistent with the one-dimensional model, suggesting that it was advected into the region.

Precipitation has been neglected in this analysis. The differences between the deep mixed casts and the model runs are within uncertainties in the estimates of evaporation minus precipitation rates. Precipitation is not an output product of the BMO model, and estimates for 1988–1989 are not available. The climatology (Gorshkov 1983) suggests precipitation may reduce the evaporation minus precipitation rates to about 35% of the values used here. This would imply that the freshwater flux from fall to winter would have to be explained through other means, for example, through the advection of ice out of the area (Roach et al. 1993; Pawlowicz et al. 1995). If one assumes the precipitation climatology is correct for 1988–1989, advection of around 15 cm of ice would be sufficient. However, the climatology shows large spatial gradients in the vicinity of the array and strong seasonal variability in evaporation rates. Better estimates of evaporation and precipitation rates are necessary to draw definite conclusions about the salinity forcing beyond this uncertainty.

These results show that a simple, one-dimensional model, with surface cooling, ice formation, and evaporation, can account for the modification of depth-averaged water properties from autumn to the deep mixed layers observed in winter. This result is not inconsistent with Fig. 3 in GSP Group (1990) but provides a different explanation for the source of salt in the mixed layer. The GSP Group (1990) suggests the extra salinity in the mixed layer is due to about 50 cm of ice growth but neglects any net evaporation minus precipitation and neglects salinity fluxes from below the level from which they integrate (1000 m). When integrating from a depth below the deep mixed layer in the central gyre, the increase in salinity from fall to winter is marginal and can be explained by reasonable evaporation minus precipitation rates.

The model output can also be compared with the observed temperature and salinity changes in the chimney, and in particular, with the evolution of the mixed layer depth. For this comparison, the model was initialized using the average of the two autumn casts shown in Fig. 11. The model mixed layer potential temperature profiles from the end of January to the time at
Fig. 12. Potential temperature profiles from model calculations for comparison with observed profiles (Fig. 7). Potential temperature profiles from the six winter GSP and SIZEX hydrographic casts within 40 km of the central mooring are superimposed at the approximate times they were taken (between 8 and 16 March). On these days model mixed layer depths are marked by black dots. Above are ice thickness output by the model and the latent (- -) and total (-) BMO surface heat fluxes used to drive the model.
which the mixed layer penetrates through 2000 m, on 22 March, are shown in Fig. 12. The model was initialized 56 days before the first day of the time period shown. The model is sensitive to the initialization profile; however, using either of the two autumn casts or their average gives mixed layer deepening to about 1200 m around the time of the winter CTD casts.

The model evolution (Fig. 12) is surprisingly similar to the observed temperature evolution of the chimney (Fig. 7). Using the −1.2°C isotherm as an estimate of the mixed layer depth in the chimney, comparable mixed layer deepening occurs in both the model and the observations during the storms on 1 and 14 February. Between these events, however, and later in the year, there are substantial differences. The observed mixed layer gradually deepens between events, whereas the model mixed layer depth does not change. Presumably other physics, such as penetrative convection, wind stirring, and shear-induced turbulence, all of which are excluded from the model, are important during these times.

The potential temperatures from the six GSP and SIZEX winter casts are superimposed in Fig. 12 at the times they were taken, for comparison with the model predictions. (As mixed layer depths are essentially the same for both temperature and salinity in the hydrographic measurements, the modeled salinity field is in comparable agreement with the hydrography.) The variability in mixed layer depths between CTD casts emphasizes the uncertainties in estimating the large-scale mixed layer depth from a few CTDs. Comparison with the inverse solutions indicates that the cast with the shallowest mixed layer has been advected into the area and that the cast with the deepest mixed layer represents an isolated plume, as discussed previously. The remaining casts are in rather better agreement with the model than might be expected.

This result differs from the model results of Pawlowicz et al. (1995), who used a quasi-static model introduced by Killworth (1979). Killworth’s model sets boundaries on mixed layer depth as a function of total surface heat flux. The minimum mixed layer depth occurs in the “rapid” limit where sensible heat fluxes dominate, and the minimum amount of ice is formed for a given amount of heat lost. Pawlowicz et al. (1995) show that the “rapid” limit does a poor job of reproducing the wintertime casts when initialized from the summer hydrography and assuming no net evaporation minus precipitation, giving mixed layers only 250 m deep in the central gyre in March 1989. Increasing salinity by removing about 50 cm of ice through advection and assuming a 10% increase in surface heat flux makes the “rapid” limit model more closely match the observed profiles in the chimney region. There is another limit to Killworth’s model, however: the “insulating” limit in which latent heat fluxes dominate and surface heat flux is used to form ice, raising ocean salinity through brine rejection. This limit gives a maximum bound on mixed layer depth for a given heat loss. Killworth notes that the actual behavior can be expected to fall somewhere between these extremes. By using a time-dependent model that predicts a mixed layer evolution between these two limits and by assuming a net evaporation minus precipitation of 0.002 m/day, as we have done here, observed mixed layer profiles in the central gyre can be explained without the need to invoke the advection of ice.

5. Potential energy budget

Previous field work in the North Pacific Ocean related the potential energy change associated with mixed layer deepening to the production of turbulent kinetic energy associated with surface forcing. Davis et al. (1981) found that the work necessary to entrain denser fluid into the deepening mixed layer can be linked to the production of turbulent kinetic energy (TKE) through wind stirring and through the interaction of entrainment stress with vertical shears. Paduan and de Szoeke (1986) concluded that potential energy changes in a shallow mixed layer are dominated by wind forcing, even during periods of strong cooling. It is expected, however, that as the mixed layer deepens, buoyancy forcing will dominate changes in potential energy. Both previous field work (Davis et al. 1981; Paduan and de Szoeke 1986) and laboratory work (Deardorff and Yoon 1984) have provided parameterizations for turbulent kinetic energy terms. Following the notation used in Davis et al. (1981) and Paduan and de Szoeke (1986), the potential energy of the mixed layer is expressed as

\[ P = g \int_{-h}^{0} (z + h) \rho dz, \]

(2)

where \( h \) is the mixed layer depth. Change in potential energy over time can be written

\[ \partial_t P = A + B + G_0 + S - E, \]

(3.1)

where

\[ A = -g \int_{-h}^{0} (z + h)(u \cdot \nabla \rho + w \partial_z \rho) dz \]

(3.2)

\[ B = hB_0 \]

(3.3)

\[ G_0 = -w' (p' + \rho(u'w' + w'^2)/2) \big|_{z=0} \]

(3.4)

\[ S = -\int_{-h}^{0} \rho u' w' \partial_z udz \]

(3.5)

The simple one-dimensional mixed layer model used in the previous section neglected mechanical forcing entirely and relied solely on buoyancy forcing.
Here \( B_0 \) is the net flux of buoyancy at the surface, expressed as the sum of temperature and salinity components,

\[
B_0 = -g \omega' \rho' |_{z=0} = B_0^T + B_0^S
\]

\[ E = \int_{-h}^{0} \rho c dz. \tag{3.6} \]

To evaluate \( \partial_z P \) from the temperature data requires that the corresponding salinity profiles be constructed. In previous works, salinity effects on density were either insignificant and neglected (Davis et al. 1981) or else of secondary importance, following the solution of Stommel.

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**Observations and Discussion**

Observations during the deep convective time period suggest that there are indeed small-scale plume features within the larger-scale chimney, which is consistent with rotationally modified convection rather than uniform mixed layer deepening. A complete analysis that accounts for rotationally modified convection is beyond the scope of this paper. We speculate that the most important term to change would be the parameterization of the convective dissipation term in (5), as the \( u^+ \) term is smaller and the production term \( B \) comes from a surface boundary condition. Jones and Marshall (1993) found from numerical simulations in an unstratified ocean, however, that within the regime where rotation is important the velocity scale within the plumes remains close to three-dimensional. Jones and Marshall’s (1993) scaling arguments suggest dissipation in the presence of rotation is also proportional to the surface buoyancy forcing, as was assumed in (5) above. To the extent that the flow is not fully three-dimensional, the constraining effects of rotation may allow less of the convective potential energy supply to be

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**6. Chimney Breakup**

After about 25 March, a one-dimensional analysis is clearly inadequate to describe subsequent developments (Fig. 10, bottom panel), and the structure of the chimney becomes more complicated. The chimney warms and losses potential energy either through radiation or advection. Figure 7 clearly shows the sporadic appearance of warmer waters suggestive of Arctic Intermediate Water, which could only be advected in from outside the chimney region.
turbulence become important when the mixed layer reaches a scale depth $l_{rot}$, which is a function of the buoyancy forcing and the Coriolis parameter only. Mixed layer depths exceed $l_{rot}$ during this time period, indicating that rotation should influence mixed layer turbulence (Fig. 13c), while the parameterizations given above neglect rotational effects. The moored observations (Schott et al. 1993) and the hydrographic
observations during the deep convective time period suggest that there are indeed small-scale plume features within the larger-scale chimney, which is consistent with rotationally modified convection rather than uniform mixed layer deepening. A complete analysis that accounts for rotationally modified convection is beyond the scope of this paper. We speculate that the most important term to change would be the parameterization of the convective dissipation term in (5), as the $u^3$ term is smaller and the production term $B$ comes from a surface boundary condition. Jones and Marshall (1993) found from numerical simulations in an unstratified ocean, however, that within the regime where rotation is important the velocity scale within the plumes remains close to three-dimensional. Jones and Marshall’s (1993) scaling arguments suggest dissipation in the presence of rotation is also proportional to the surface buoyancy forcing, as was assumed in (5) above. To the extent that the flow is not fully three dimensional, the constraining effects of rotation may allow less of the convective potential energy supply to be available for vertical entrainment, which could account for some of the discrepancy during the deep mixed layer time period in Fig. 13b.

A potentially important source of convective TKE energy that has been ignored is that associated with brine rejection from ice formation. While it is not possible to determine ice growth rates from SSM/I observations, the hints of low ice concentrations during times of active convection leave open the role ice might play, if any, with respect to convection. The time series in Fig. 13b indicate that, at least within the accuracy of our measurements and to the extent that three-dimensional turbulence applies, extra-turbulent kinetic energy production associated with brine rejection is unnecessary to explain the observed changes in potential energy.

6. Chimney breakup

After about 25 March, a one-dimensional analysis is clearly inadequate to describe subsequent developments (Fig. 10, bottom panel), and the structure of the chimney becomes more complicated. The chimney warms and loses potential energy either through radiation or advection. Figure 7 clearly shows the sporadic appearance of warmer waters suggestive of Arctic Intermediate Water, which could only be advected in from outside the chimney region.

The low-passed current structure at the central mooring during this time shows significantly different characteristics as well (Fig. 14). The central mooring is close to the perimeter of the chimney. The most notable feature is the sudden increase in variance at the time the chimney region loses the colder deep convected waters.
A widely discussed mechanism for chimney breakup is baroclinic instability (Killworth 1976; Gascard 1978; Gascard and Clarke 1983; Hermann and Owens 1993; Saunders 1973), in which a chimney breaks up into smaller eddies and disperses. The increases in both temperature structure and velocity variance at the end of the strong surface heat loss after about 25 March seem qualitatively consistent with breakup by baroclinic instability. Measured and predicted timescales are compared qualitatively below.

Hermann and Owens (1993) use analytical methods, in addition to constructing a hydrostatic, primitive equation model, to look at the characteristics of chimney adjustment after the surface forcing ends. This is applicable to the chimney observed here given the drop in surface forcing after 15 March. They find that narrower chimneys tend to lose a greater fraction of their energy through inertial waves, while a slower advection process is the primary means of energy loss for wider chimneys. Their “wide” chimney simulations (8.5 Rossby radii, or 25 km in diameter) break into three smaller eddies with a timescale of about 8 days, compared to the timescale 3–6 days in the observations.

Legg and Marshall (1993), instead, consider the chimney as a collection of small-scale plumes. They incorporate results from the nonhydrostatic numerical model presented in Brugge et al. (1991) and use a two-layer heton model first developed by Hogg and Stommel (1985). Hetons are stacked pairs of point-vortices, with a positive (negative) vorticity anomaly in the upper (lower) layer. Each heton is used to represent a single geostrophically adjusted plume. They find that a collection of hetons introduced into a chimney-scale region will interact and set up a rim current that dynamically confines the vortices. When enough hetons are present, however, clumps of vortices on the Rossby deformation scale have enough energy to penetrate the rim current, carrying heat through the sides of the chimney and balancing the net surface forcing. Using typical numerical values for Mediterranean open-ocean con-
vection, the chimney breakup timescale is faster than that found in Hermann and Owens (1993), around one to two days depending on the strength and spatial distribution of hetons added to the chimney and the strength of atmospheric forcing. Although the temporal (3 day) and spatial smoothing in our inversions make the data inadequate to differentiate between these simulations, the model time evolutions are comparable to the chimney evolution that we observe. From Fig. 3a, the initial cooling between 500 and 1000 m is most intense and strongly confined on 19 and 22 March and then weakens by 28 March as the forcing decreases. The observed 3–6 day breakup falls between the 1–2 day timescale in the heton model, and the 8-day timescale from the Hermann and Owens (1993) ‘‘wide’’ chimney results.

7. Discussion

a. Why did the chimney occur where it did?

Previous hydrographic observations in other seas, such as the Weddell Sea (Killworth 1979), suggest that large areas should convect given the observed atmospheric forcing, yet convection is confined to small regions. Some physical process must be invoked to preferentially modify a region for later convection. Potential preconditioning agents include topography in the Mediterranean, Labrador, and Weddell Seas (Hogg 1973; Gascard and Clarke 1983; Gordon 1978); cyclonic gyre-scale circulation in the Greenland and Mediterranean Seas (Swallow and Caston 1973; Killworth 1979); and cyclonic eddies in the Boreas Basin (Johannessen et al. 1991). Ice cover could also precondition an area by brine rejection (Rudels 1990). Favorable winds could induce ice-edge upwelling (Häkkinen 1987).

To see whether the chimney region itself has a ‘‘preconditioned’’ temperature signal that sets it apart from typical gyre waters even before the ice appears, we compare the average temperature of the chimney area over the seasonal cycle to the gyre average (Fig. 15). The chimney region is significantly colder than the gyre average from the end of October through the winter. The warmer, saltier type of Arctic Intermediate Water, which lies at intermediate depths and is believed to play a key role in the deep water formation process (e.g., Killworth 1983; Carmack and Aagaard 1973; Worces-ter et al. 1993; Pawlowicz et al. 1995; Morawitz et al. 1996), does not exist homogeneously over the gyre at the beginning of the winter season. Instead there are holes in the AIW coverage that result in relatively colder regions with lower stratification (see Fig. 13 of Morawitz et al. 1996).

The deep convective process we observe appears to be sensitively linked to the amount of A IW present from autumn through spring. As has been discussed, the salinity in these waters is important to destabilizing the water column when this water mass is brought to the surface and cooled. However, too much A IW overstratifies the water column and prevents a deep mixed layer from developing. In this case, surface mixing cannot penetrate the intermediate layer density gradient and the cooling is confined near the surface rather than being mixed into progressively deeper waters. The surface waters eventually freeze, and the subsequent insulating ice cover reduces the thermal forcing at the surface.

One other region of the gyre to the northeast of the central mooring also appears to be ‘‘preconditioned’’ by lacking the warmer Arctic Intermediate Water. However, warmer water apparently advects into this area in early March, restratifying the water column and preventing the chimney from maturing. We speculate that the circulation of Arctic Intermediate Water responsible for forming these holes in A IW coverage could be what effectively preconditioned deep convecting regions within the Greenland gyre. A better understanding of the circulation within the Greenland Sea is needed to explain why these features appear where and when they do.

b. What was the plume concentration in the chimney that formed in the Greenland Sea during winter 1988/89?

Despite relatively heavy CTD coverage close to the time the chimney matured, only one hydrographic cast showed a deep mixed layer to 1500 m, presumably capturing a plume. We use the temperature fields from the combined inversions, and that single cast, to estimate the plume concentration within the chimney for comparison to other estimates. The plume is ~0.12°C colder between 1000 and 1500 m than the other hydrographic casts in the chimney (Fig. 4), while the chimney is on average 0.035°C colder than surrounding profiles over this depth range. Assuming each plume has a similar temperature signal, and that all the cooling over this depth range is due to plumes, the volume fraction occupied by plumes is 0.035/0.15, which is about 20% of the chimney volume. This is consistent with the estimate made by Pawlowicz et al. (1995) for the 4 to 6 path, a significant fraction of which passed through the chimney.

c. How much deep water formed?

We estimate the amount of deep water formed from the average temperature change between 1000 m and 2000 m over the entire array from autumn 1988 until the end of the convective season: approximately 0.025 ± 0.006°C. Assuming cooling occurs in the form of plumes and that the deep-mixed CTD cast is representative of plume temperatures, cooling occurred over (0.025 ± 0.006)/0.15 = 0.17 ± 0.04 of the 2.18 × 10^13 m^3 array volume. Therefore, 0.17 ± 0.04·2.18
\[ \times 10^{13} \text{ m}^3 = 3.71 \times 10^{12} \pm 0.87 \times 10^{12} \text{ m}^3 \] of new water formed between 1000 and 2000 m during 1988–1989, so that the annual production rate is about 0.12 \pm 0.03 Sv (Sv = 10^6 \text{ m}^3 \text{s}^{-1}).

From tracer studies in the Greenland Sea, Rhein (1991) and Schloesser et al. (1991) conclude that since 1982–1983 the ventilation rate of the deep waters has decreased to about 10% of its value in the 1970s. Rhein (1991), using CFCs, finds a range of 0.085–0.38 Sv for the annual-average deep-water production rate of new deep water below 1500 m, depending on model assumptions. Schloesser et al. (1991), using a wider range of tracers, estimates an annual deep-water formation rate in the 1980s of “about 0.1 Sverdrups.”

Although the production rate estimates made here are for a shallower depth range, it is generally consistent with the tracer estimates suggesting 1988–1989 was a typical year for Greenland Sea Deep Water formation during the 1980s and early 1990s (Rhein 1994). A significant fraction of deep Greenland Sea waters can evidently be formed by convective chimneys.

**Acknowledgments.** Walter Munk was instrumental in the design and conduct of the experiment. His advice on the analysis is greatly appreciated. We also wish to thank J. Swift and K. Aagaard for helpful discussions. J. Swift was instrumental in compiling the 1988–1989 hydrographic dataset. The following graciously provided data for the inversions: K. Aagaard, G. Budeus, O. M. Johannessen, E. Fahrbach, J. Meincke, F. Schott, and J. Swift. R. Schuchman provided the SSM/I ice data. Credit for the success of the tomographic experiment belongs largely to the dedicated personnel who designed, fabricated, tested, and fielded the tomographic instruments: S. Abbott, J. Bouthilette, P. Boutin, K. Hardy, D. Horwitt, J. Kemp, S. Liberatore, D. Peckham, and R. Truesdale. B. Ma and A. Newhall assisted with programming for the data analysis. B. Betts prepared the illustrations. The Greenland Sea Tomography Experiment was supported by the National Sea Foundation (Grants DPP9102061 and DPP8518643) and the Office of Naval Research (ONR N00014-86-G-0150).

**APPENDIX**

**Brief Details of the Combined Inversions**

Details of the combined inversions are in Morawitz et al. (1996), and we only provide a brief summary here.

A statistical parameterization (or “model”) of the sound speed (temperature) field is made consisting of horizontal and vertical basis functions. The vertical basis set is constructed from a smooth data covariance matrix based on all available historical and 1988–1989 hydrographic data. The covariance is normalized before being decomposed. The 12 largest renormalized eigenvectors that comprise the vertical basis can account for 90% of the variance at every depth. The horizontal model is based on cross-correlations between thermistor time series at the same depth on different moorings. The horizontal basis set is spectral. Transformed into physical space, the horizontal model covariance is approximately Gaussian with an e-folding scale of 40 km. In the model, the predominant source of noise in the point measurements is short-scale (<40 km) horizontal energy, comprising 20% of the total variance.

The inversions are done for sound speed and then converted to temperature. The acoustic travel times are expressed as perturbations to travel times computed for a reference state \( C_0(z) \). Expanding in the model basis,

\[
\Delta T_i = -\sum_j \sum_{kl} a_{ijkl} \int_{r_i} \frac{ds}{C_0^2(z)} F_j(z) G_{kl}(x, y) + \int_{r_i} \frac{u \cdot \tau}{C_0^2(z)} + r_i^T, \quad (A1)
\]

where \( \Delta T_i \) is the measured travel time minus the travel time through the reference state, \( a_{ijkl} \) are the model parameters, \( \Gamma_i \) is the acoustic ray path, \( u \) is the current velocity, \( \tau \) is the unit vector tangent to the ray path \( \Gamma_i \), and \( r_i^T \) is the error. The second term on the right-hand side of (A1) is small and is either included in the error term or cancels when sum travel times are used.

For the point measurements, the measured perturbation field can be written

\[
\Delta C(x, y, z) = C(x, y, z) - C_0(z) = \sum_j F_j(z) \sum_{kl} a_{ijkl} G_{kl}(x, y) + r^C. \quad (A2)
\]

To convert thermistor measurements to sound speed when salinity data are not available, a fixed salinity of 34.895 psu is assumed.

Equations (A1) and (A2) can be combined into a single matrix equation, and the model parameters \( a_{ijkl} \) found using standard weighted least squares inverse methods. The solution can then be written in terms of the estimated model parameters \( \hat{a} \)

\[
\Delta \hat{C}(x, y, z) = \sum_{jkl(m)} \hat{a}_{jkl(m)} F_j(z) G_{kl(m)}(x, y). \quad (A3)
\]

The sound speed field is converted to potential temperature assuming the same fixed salinity as above. Errors due to expected salinity variability can be as large as 0.050°C near the surface but are insignificant over the rest of the water column. The uncertainty of the solution is estimated from the inverse formalism in the standard way.

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