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News from the WOCE IPO

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Research ships, tall ships and AUVs

Two recent events have reinforced in my mind a fact of which we should not lose sight – making observations at sea is still central to our understanding of the oceans but is at times very difficult.

In February Penny Holliday, known to many of you as the person in the WOCE IPO with responsibility for data issues, led her first cruise. She and her team set off to occupy a hydrographic section from the UK to Iceland including the often repeated "Ellett" Line from the UK to Rockall. February is never the best time of year to work in the N. Atlantic but one objective was to observe end-of-winter conditions. They endured winds of 30 m/s and waves approaching 30 m in height and despite their best efforts they were only able to occupy shelf stations and the Ellett Line.

In mid-April almost 20 large old sailing ships set off from Southampton on a Tall Ships race to Cadiz, Bermuda, Newport RI, Halifax NS, ending in Amsterdam. They were an impressive sight and attracted thousands of visitors. The Southampton Oceanography Centre mounted a series of exhibits (including one about what WOCE and CLIVAR are doing to better understand climate variability and change). The displays were both backward-looking (to the days of sail when HMS Challenger made the first global ocean survey) and forward-looking (to the use of AUVs and profiling floats). The day is not far away when these new techniques and satellite and maybe acoustic remote sensing will provide much of our information about the oceans. Even then, people like Penny and her team will still have to endure considerable discomfort to collect the complementary ship-board ocean observations.

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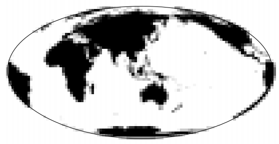
The dynamics of ocean heat transport variability



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One of the principal goals of WOCE is to understand the global large-scale ocean heat transport and its variability over a broad range of time and length scales (World Climate Research Programme, 1986). The north–south heat transport is the prime manifestation of the ocean's role in global climate, but understanding of its variability has been fragmentary owing to uncertainties in observational analyses, limitations in models, and the lack of a convincing mechanism. With the completion of the observational phase of WOCE, more hydrographic sections are now available and better estimates of the global time-mean ocean heat transport are possible. In general, it can be said that the sign and magnitude of the ocean heat transport are known over the global ocean, and that quantifiable error estimates can be made. Hence, it is timely to consider the time-dependent nature of ocean heat transport, the nature and magnitude of which are not well known, with extant estimates differing not only in magnitude but in sign as well.

In this note we touch upon some of the questions concerning the ocean's role in climate, by addressing temporal variability in the ocean's transport of heat and, in particular, the global nature of the relevant ocean dynamics (this article summarises some results contained in a more detailed paper Jayne and Marotzke, 2000). An ocean general circulation model (the Parallel Ocean Climate Model, POCM, Semtner and Chervin, 1988, 1992; Stammer et al., 1996; McClean et al., 1997) is used to understand the ocean's response to the seasonally varying wind stress. The numerical simulation output from run 4_B of the POCM, is used to calculate ocean mass and heat transport at three day intervals. The POCM is a primitive-equation, level model configured for the global ocean between 75°S and 65°N, with realistic topography. The heat transport from the model was calculated at 3-day intervals and the time-mean heat transport was then removed. The results are presented in a Hovmöller diagram (Fig. 1, page 22) of the heat transport anomaly (from the time-mean) as a function of latitude and time for the World Ocean, and the annual cycle for the ocean basins is summarised in Fig. 2.

Globally, the cross-equatorial, seasonal heat transport fluctuations are close to $\pm 4.5 \times 10^{15}$ watts, the same amplitude as the seasonal, cross-equatorial atmospheric energy transport. The seasonal cycle of ocean heat transport is directed from the summer hemisphere to the winter hemisphere (i.e. northward in boreal winter and southward in austral winter), in phase with the annual cycle of total energy transport by the atmosphere's Hadley cell (Peixoto and Oort, 1992). At 7°N, the Atlantic and Pacific Oceans have their maximum amplitude in the seasonal cycle of 1 PW and 3 PW, respectively. The Indian Ocean has its

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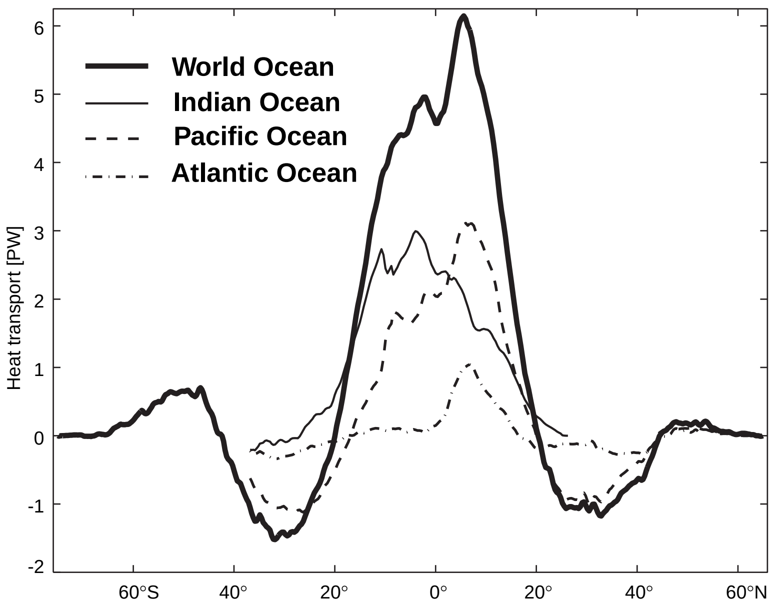

Figure 2. Annual cycle of heat transport defined as the difference between averaged January and averaged July values for the World Ocean (heavy solid line), the Indian Ocean (thin solid line), the Pacific Ocean (dashed line) and the Atlantic Ocean (dashed–dotted line).

maximum peak-to-peak seasonal cycle of 2.6 PW at 5°S. Globally, the variability is concentrated within 20° of the equator and dominated by the annual cycle, with higher-frequency variability superimposed on it.

To explain this large variability, we build upon the approach put forth by Bryan (1982) which we expand by linking it to the theoretical work of Willebrand et al. (1980) and Schopf (1980). In summary, the majority of the heat transport variability can be ascribed to the "Ekman heat transport". Introduced by Bryan (1962), it was expanded upon by Bryan (1982) in his analysis of early modelling work, and then used with observational data by Kraus and Levitus (1986) and later by Levitus (1987), Adamec et al. (1993) and Ghirardelli et al. (1995). The definition of the Ekman heat transport given by Kraus and Levitus (1986)in the following equation:

$$H(t) = -\int \rho_0 c_p \frac{\tau_\lambda}{f\rho_0}\{T_{Ek} - [\theta]\} a \cos\phi \, \mathrm{d}\lambda$$

where c_p is the specific heat of seawater, f is the Coriolis parameter, $T_{Ek}(x)$ is the temperature of the surface Ekman layer, $[\theta]$ is the zonal and depth-averaged potential temperature, $\tau_\lambda(x)$ is the zonal wind stress, and a is the radius of the earth.

Simply put, this equation expresses the heat transport as the integral of the Ekman mass flux times the difference between the Ekman layer temperature and the section averaged potential temperature. It implies that for any given section the mass transport in the Ekman layer is compensated by a return flow distributed uniformly across the depth and zonal extent of the section. Indeed, Kraus and Levitus (1986) and succeeding writers, such as Levitus (1987), Adamec et al. (1993), and Ghirardelli et al. (1995), all assumed that this equation held over all timescales. Adamec et al. (1993) and Ghirardelli et al. (1995) also made the further assumption the local Ekman transport was returned locally at the same longitude, but this is not necessary. In these previous works both the time-varying and the time-mean flows have been lumped together. The assumption that the return flow for the time-varying Ekman transport is deep and barotropic is supported by theory (Willebrand et al., 1980; Schopf, 1980; Jayne and Marotzke, 2000) and modelling studies (Bryan, 1982; Böning and Herrmann, 1994; Lee and Marotzke, 1998). However, Klinger and Marotzke (2000) have shown that the time-mean Ekman mass transport is returned at shallow depths, and relatively warm temperatures, within the main thermocline. Therefore, the heat transport by the time-mean Ekman flow would not be well represented by the assumption that its return flow was barotropic.

To compare the heat transport variability in the full model to the Ekman heat transport, a simple estimate from data can be made of the annual cycle based on the climatology of ocean temperature (Levitus et al., 1994) and monthly-averaged wind stress fields derived from the same fields used to force the POCM simulation. The results are shown in Fig. 3. Overall, the agreement between the heat transport variability in POCM and this simple calculation are excellent, and supports our conclusion that the time-dependent ocean heat transport is essentially given by the time-varying part of the Ekman heat transport.

In summary, the temporal variability in the ocean heat transport is dominated by the Ekman heat transport. The deep compensating return flows have little shear associated with them and therefore would not affect estimates of the heat transport from hydrographic surveys in any of the ocean basins, in agreement with the findings for the Atlantic Ocean of Böning and Herrmann (1994). As for other sources of variability, away from the tropics, the heat transport variability associated with the barotropic gyre and baroclinic circulations, are much weaker than the Ekman variability, and can amount to a 0.2–0.4 PW variance in the heat transport measured by a one-time hydrographic survey. Hence estimates of the time-mean heat transport made from one-time hydrographic surveys using the method of Hall and Bryden (1982) are fundamentally sound. One of the goals of future observations and modelling efforts should be to understand smaller, but perhaps locally significant, baroclinic heat transport variations, as well as longer time scale variability.

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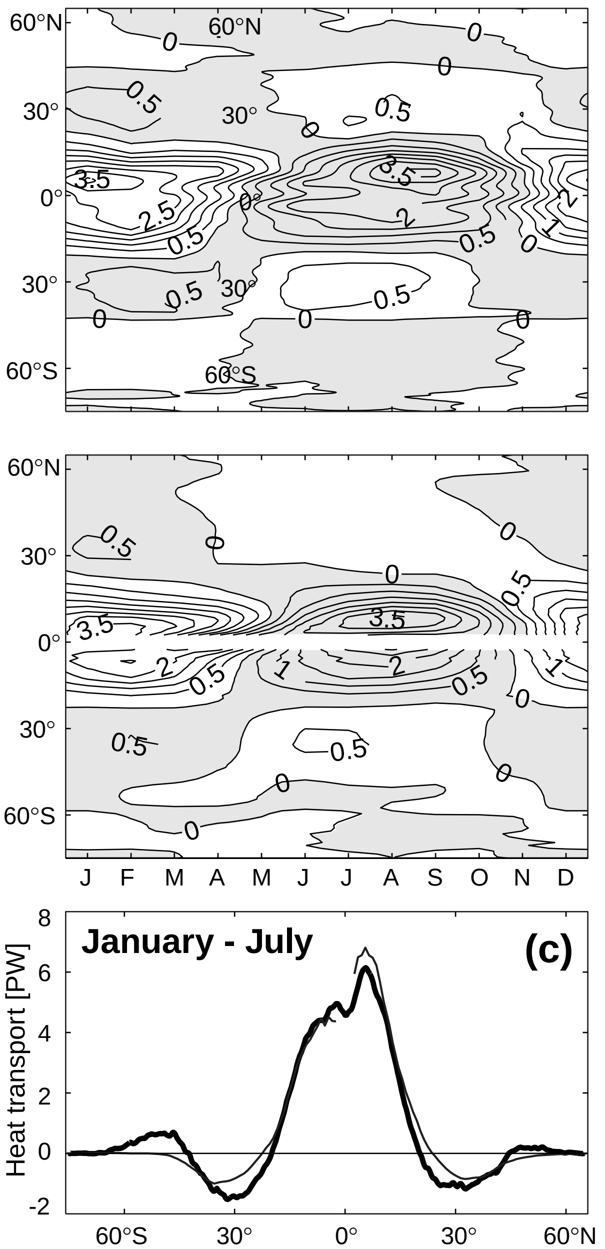

Figure 3. (a) Annual cycle of heat transport from POCM. (b) Annual cycle of Ekman heat transport from wind stress and temperature anomalies for the World Ocean using the Levitus et al. (1994) climatology and (1). (c) The annual cycle (January–July) from the POCM (heavy line) versus climatology (thin line) for the World Ocean. Contour interval for (a) and (b) is 0.5 PW.

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