# 4.7

# Interocean Exchange

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## 4.7.1 Interocean Links

Earth's climate, responding to the different thermodynamic properties of the land and ocean surfaces, is sensitive to the continental configuration and distribution of mountain ranges. This is clearly seen in the pattern of mean annual and seasonal range in such climate parameters as temperature and humidity and in the quasistationary patterns of atmospheric circulation, from small-scale sea breezes to planetary waves. Presumably because of the ocean–land configuration, each ocean basin is exposed to different atmospheric forcing, taking on correspondingly distinct property and circulation characteristics (**Figs 4.7.1a,b,** see Plate XX). These in turn provide feedback to the climate system through their effect on Sea Surface Temperature (SST) distribution, heat and freshwater fluxes and ocean overturning. Similarity between oceans is inhibited by their varied degrees of isolation from one another, and thus the coupled ocean–atmosphere system is influenced by the efficiency of interocean exchanges that link the ocean basins. More efficient interocean exchange leads to reduced contrast between the oceans, with each ocean closely resembling its neighbour. Interocean exchange would be expected to be balanced mainly by ocean circulation on approximately a horizontal plane. Less efficient exchange is expected to produce oceans that contrast sharply with each other, a condition more apt to induce stronger global reaching, overturning thermohaline circulation. Equilibrium states between patterns of interocean exchange and circulation on the horizontal and vertical planes may be expected for specific continental configurations.

### *4.7.1.1 Antarctic Circumpolar Current*

The oceans of the southern hemisphere vary less from each other than do the northern hemisphere oceans. Isolation and differences grow with distance from the rapid interocean exchange afforded by the Antarctic Circumpolar Current (ACC; for detailed information of the ACC, see Rintoul *et al.*, Chapter 4.6). The ACC is the giant of interocean exchange, carrying about 134 Sv (Sverdrup, $1\,\mathrm{Sv} = 10^6\,\mathrm{m}^3\,\mathrm{s}^{-1}$) of polar and subpolar water masses from west to east through the Drake Passage (Nowlin and Klinck, 1985). Variations in ACC transport through the Drake Passage amount to 20% of the mean. ACC transport is enhanced south of Australia by the Indonesian Throughflow, of about 10 Sv.

The ACC is mainly a zonally flowing current, but large quasistationary waves in the ACC, guided by bottom topography, lead to a latitudinal swing of approximately 1200 km (furthest north in the Atlantic; furthest south in the southwest Pacific – Gordon *et al.*, 1978; Orsi *et al.*, 1995). The equatorially flowing Malvinas Current, which may be considered as a branch of the ACC, carries subpolar waters well to the north, to the separation of the Brazil Current near 38°S. Transient waves in the ACC (Antarctic Circumpolar Wave; White and Peterson, 1996) may link the ACC with sea ice distribution and larger-scale climate variability (Yuan and Martinson, 2000). Mesoscale eddies within the ACC provide oceanic meridional heat and freshwater fluxes to balance much of the ocean–atmosphere exchange south of the ACC.

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The Southern Ocean meridional overturning is induced by the large-scale wind and by buoyancy forcing of the very cold conditions along the margins of Antarctica. Deep water (potential temperature 1–2°C, salinity 34.7–34.9) with characteristics developed in more northern latitudes, upwells around Antarctica to be converted into Antarctic Intermediate Water (AAIW; 3–5°C, 34.2–34.4) and Antarctic Bottom Water (AABW; $< -1$°C, 34.65–34.75). Southern Ocean overturning induces large upward heat flux, limiting sea ice thickness (Gordon and Huber, 1990). Southern Ocean overturning is most effective in projecting Southern Ocean water mass properties to the global scale (see Sloyan and Rintoul, 2000).

An interesting aspect of the ACC relevant to interocean heat and freshwater fluxes, the theme of this chapter, which may contribute to the saltiness of the Atlantic Ocean, was pointed out by Gordon and Piola (1983), who inspected the salinity change of the subpolar surface water within the ACC *en route* across the South Atlantic. They found that the observed freshening equals the net evaporation minus precipitation of the subtropical South Atlantic. The water vapour derived from the subtropics is carried southward, to balance excess precipitation over the ACC (see Fig. 12.10 of Peixoto and Oort, 1992). The ACC exports the fresh water across the Indian Ocean into the Pacific Ocean. This situation arises because the poleward extension of South America and the Andes limits freshwater transfer from the Pacific to the Atlantic by both ocean and atmosphere. There is no such block provided by Africa, which does not reach the position of the maximum westerlies. The same may be true of the Indian sector, as southern Australia is also well north of the maximum westerlies. Extraction of fresh water from the South Atlantic increases the salinity of the South Atlantic subtropical water, which has an extensive region of surface salinity of greater than 36.4, an attribute lacking in the other southern hemisphere subtropical gyres (**Fig. 4.7.1b,** see Plate XX). Saline South Atlantic surface water eventually spreads northward within the North Brazil current and may be a factor in the saltiness of the North Atlantic Ocean.

#### *4.7.1.2 Northern oceans*

The great ocean embayments of the northern hemisphere vary markedly from each other. Surplus precipitation and runoff over evaporation in the North Pacific induces low surface salinity with a highly stable halocline inhibiting deep-reaching convection. The lack of deep convection in the subpolar North Pacific and associated meridional overturning circulation limits poleward spreading of warm low-latitude waters, which further suppresses evaporation (Warren, 1983). The North Atlantic Ocean, with excess evaporation over precipitation and runoff, forms a saline surface layer prone to deep convection and the formation of North Atlantic Deep Water (NADW). Sinking of surface water associated with NADW formation draws compensatory warmer surface water poleward. Warmer SST in the western subtropical North Atlantic (relative to that of the North Pacific, **Fig. 4.7.1a**) may encourage further evaporation, invigorating Atlantic meridional overturning circulation. Evaporation of the subtropical North Atlantic and net precipitation over the equatorial Pacific are coupled by westward water vapour flux (about 0.3 Sv; Zaucker and Broecker, 1992) across the Isthmus of Panama, thought to be a major force behind NADW formation (Zaucker *et al*., 1994).

The Pacific halocline abruptly developed, along with intensification of northern hemisphere glaciation, about 2.73 million years ago (Haug *et al*., 1999). It is likely that this coincided with severing of the direct oceanic link between the North Pacific and North Atlantic Oceans with the rise of the Isthmus of Panama and, by inference, the development of a saline North Atlantic, conductive to NADW formation. Thus a change in interocean exchange 2.73 million years ago forced a different pattern of horizontal and vertical circulation that remains active in today's ocean.

The Indian Ocean north of 10°S is exposed to a strongly monsoonal climate. Convection into the thermocline and intermediate levels occurs within the evaporative, saline Arabian and Red Seas. Enormous influx of fresh water produces stratification reminiscent of an estuarine environment within the Bay of Bengal. Exchange of fresh water between the Arabian Sea and the Bay of Bengal may be viewed as a regional analogue to the Atlantic–Pacific global-scale system.

#### *4.7.1.3 The global chain of interocean thermohaline links*

Interocean exchange is suspected as being an important part of the present-day global thermohaline circulation, particularly that which is in response to NADW formation (Gordon, 1986, 1996a,b; Broecker, 1991; Rintoul, 1991; Gordon *et al.*, 1992; Schmitz, 1995; MacDonald and Wunsch, 1996). The reader is directed to the excellent two-volume report of Schmitz (1996a,b).

NADW is exported into the Indian and Pacific Oceans by the ACC, becoming entangled in the overturning circulation of the Southern Ocean and associated AAIW and AABW formation. NADW export from the Atlantic Ocean must be balanced by import of Indian and Pacific Ocean waters within the water column shallower than NADW – but how? From the warm saline Indian Ocean thermocline and intermediate waters around the southern rim of Africa (the warm route), or from cooler, fresher subpolar Pacific water through the Drake Passage (the cold route)? Or, as more likely, both? Does the Indonesian Throughflow have anything to do with NADW formation? Do the ratio or efficiencies of these return routes vary in time? Might such variability be coupled to NADW formation and climate variability? Though there has been a plethora of papers on these subject, some favouring the warm route (a recent example is that of Holfort and Siedler, 2000) and others the cold route (Rintoul, 1991; Schlitzer, 1996), definitive answers as to the climate importance of interocean exchange are still evolving.

MacDonald and Wunsch (1996) investigated the nature of the global thermohaline circulation pattern using a set of 23 (mostly WOCE) sections and the statistical guidelines of the inverse box model approach. Within the limits of the non-synoptic data set, two independent large-scale, global integrated circulation cells emerge. The Atlantic–Southern Ocean cell carries NADW into the Southern Ocean, where it is integrated into the Southern Ocean overturning cell and spreads into the Indian and Pacific Oceans. The second cell is confined more to a horizontal plane, linking the Pacific and Indian Oceans by westward flow within the Indonesian Seas and eastward flow (presumably within the subpolar zone) south of Australia. The two cells are linked through the highly time-dependent Agulhas Retroflection south of Africa and through upwelling of deep waters in the Pacific Ocean (Gordon, 1996b).

Interannual to decadal SST anomalies may be transferred between ocean basins by the interocean links (e.g. the ACC; Peterson and White, 1998). Also, one can envision that SST anomalies could be generated within an ocean basin by variability in the interocean transport (e.g. variability in any of the three interocean channels discussed below: Bering Strait, Indonesian Seas and Agulhas Retroflection). Millennium-scale changes in wind and sea level associated with glacial epochs may also be expected to alter the form of the interocean exchange, which would alter global thermohaline circulation with feedback to the climate system (Seidov and Haupt, 1999).

While the ACC is by far the largest conduit for interocean exchange, water mass differences between the major ocean basins would be much larger were it not for various smaller interocean links, which may be linked into a global chain. The objective of this chapter is to present the current state of knowledge about those small but important interocean links. The regional oceanography of these areas is not developed (see Tomczak and Godfrey, 1994), but rather only those aspects directly associated with interocean fluxes. The interocean links discussed below are: the Bering Strait (**Fig. 4.7.1c**, see Plate XX) and Indonesian Seas (**Fig. 4.7.1d**, see Plate XX), which allow for the export of low-salinity North Pacific upper layer water to the Atlantic (Arctic) and Indian Oceans, respectively; and the Agulhas leakage of Indian Ocean thermocline and intermediate water into the Atlantic at the southern rim of Africa (**Fig. 4.7.1e**, see Plate XX). Before discussing these interocean links, it is worthwhile mentioning the presence of westward flow of Pacific water into the Indian Ocean immediately south of Australia.

#### *4.7.1.4 South of Australia*

Observations (Fine, 1993; Reid, 1997; Rintoul and Bullister, 1999) and models (Semtner and Chervin, 1992; Speich *et al.*, 2000) suggest flow of Pacific water into the Indian Ocean along the southern coast of Australia. While much of this may be part of a closed anticyclonic gyre in the Great Australian Bight, there is a possibility of waters from the Tasman Sea flowing into the Indian Ocean. Rintoul and Bullister (1999) find 2–3 Sv of Tasman water flowing westward south of Tasmania within the 800 and 3000 m depth interval; Speich *et al.* (2000) in their model study find 3.2 Sv of Pacific water entering the Indian Ocean in the upper 1200 m, and the model shows this water crossing the Indian Ocean to flow south through the Mozambique Channel, Agulhas Current and into the South Atlantic. Whether the westward flow immediately south of Australia is an important, overlooked element of a global-scale interocean circulation pattern, or more of a regional, Indian Ocean gyre reaching into the Tasman Sea, is an important issue to be resolved. The basic question is, 'Is the South Pacific "climate" imprinted on the Pacific to Indian flow south of Australia?'

## 4.7.2 Bering Strait

### *4.7.2.1 Introduction*

The Bering Strait, with a sill depth of 45 m within Anadry Strait (about 200 km south of the narrowest width of Bering Strait), allows cold, low-salinity surface waters to flow from the North Pacific to the Atlantic's Arctic Sea (**Fig. 4.7.1b**, see Plate XX). While the mass transport of less than 1 Sv is minor compared with other interocean flows, it is notable because of the influence of the Bering strait transport on the Arctic freshwater budget (Aagaard and Carmack, 1989, 1994; Swift *et al*., 1997). North Pacific waters entering the Arctic Sea spreads within the Arctic upper pycnocline layer into the Canada Basin. The Bering Strait water eventually is exported from the Arctic to the northern North Atlantic Ocean via the Fram Strait and within the complex channels of the Canadian northwest territory (Rudels *et al*., 1994; Jones *et al*., 1998).

### *4.7.2.2 Transport*

A 4-year time series of temperature, salinity and velocity data across the Bering Strait from 1991 to 1994 is presented by Roach *et al*. (1995). They find a mean transport of 0.83 Sv with a weekly standard deviation of 0.66 Sv. The annual cycle has a range of 1 Sv, with a maximum in summer, and a secondary maximum in January. Interannual variability of 0.5 Sv is observed. The northward transport of Pacific water is linearly linked to the meridional wind (transport $= 1.06 - 0.12\,V$, where $V =$ meridional wind in m s$^{-1}$). Roach *et al*. (1995) find that the salinity of the throughflow is near 32 in autumn and about 34 in the spring, the difference reflecting summer ice melt and winter ice formation. Niebauer (1998) shows large changes in sea-level atmosphere pressure over the North Pacific as the Aleutian low shifts zonally in response to El Niño and La Niña phases (further east during El Niño). Large interannual variability of the transport and properties of the Bering Strait Throughflow may be related to changing sea ice distribution and wind responding to the shifting Aleutian low.

### *4.7.2.3 Thermohaline fluxes*

Aagaard and Carmack (1989) stress the importance of salinity to convective overturning within the Greenland, Iceland, Norwegian and Labrador Seas. Much of the low-salinity surface water carried into these seas by the East Greenland Current is derived from the North Pacific via the Bering Strait. Investigating the freshwater budget for the Arctic Sea (defined as the region between Fram Strait and Bering Strait) relative to 34.80, the mean salinity of the Arctic Sea, they find that the Bering Strait Throughflow of 0.8 Sv supplies 1670 km$^3$ yr$^{-1}$ of fresh water to the Arctic Sea; if spread evenly over the Arctic Sea, this amounts to 18 cm yr$^{-1}$. This is the second largest source of fresh water for the Arctic Sea, the largest being river runoff (3300 km$^3$ yr$^{-1}$, about 0.13 Sv of fresh water, providing an Arctic freshwater cover of 35 cm yr$^{-1}$).

Wijffels *et al*. (1992) begin their global assessment of oceanic freshwater fluxes with the Bering Strait, using 0.8 Sv of 32.5 salinity water. In their analysis '... freshwater transport applies to that part of a seawater flux that is pure water.' As seawater is roughly 3.5% salt, the freshwater component is 96.5%. Wijffels *et al*. (1992), using the Bering Strait flux and the Baumgartner and Reichel (1975) sea–air freshwater flux values, find 0.75 Sv of fresh water enters the Arctic Sea from the Pacific. That, with the 0.18 Sv (0.05 Sv more than the Aagaard and Carmack value) of fresh water added to the Arctic by excess of runoff and precipitation minus evaporation, yields 0.93 Sv of fresh water exiting the Arctic Sea, across the latitude of Iceland.

In summer, when the Bering Strait throughflow transport is at a maximum, it injects its properties into the 50–100 m layer of the Arctic Sea, inducing a weak subsurface temperature maximum. In winter, colder and more saline Bering Strait water spreads into a deeper Arctic layer, 150–200 m, producing a temperature minimum (Tomczak and Godfrey, 1994; Rudels *et al*., 1996). Cold, low-salinity Bering Strait water contributing to the Arctic pycnocline, acts to isolate the warmer deeper water derived from the Atlantic from the ice covered surface, reducing vertical heat flux and promoting a thicker sea ice cover than found in the Southern Ocean, where the temperature and salinity profiles coincide.

The Bering Strait water, boosted by an additional 0.13–0.18 Sv of river inflow and excess precipitation, ultimately passes into the North Atlantic, where it provides the fresh water to offset the net evaporation over the Atlantic Ocean and input of salt from the Indian Ocean, conceivably playing a vital role in the susceptibility of the Atlantic Ocean to convection (Rahmstorf, 1995, 1996; Weijer *et al*., 2000a; see Section 4.7.5).

### 4.7.3 Indonesian Seas

#### *4.7.3.1 Introduction*

A pathway for more massive export of Pacific upper ocean water to an adjacent ocean than afforded by Bering Strait occurs within the Indonesian Seas (**Fig. 4.7.1d**, see Plate XX; see also Godfrey, 1996; Lukas *et al*., 1996). Large-scale observation-based studies (including inverse solutions) reveal significant Pacific export of fresh water and heat into the Indian Ocean (Piola and Gordon, 1984, 1986; Toole and Raymer, 1985; Wijffels *et al*., 1992; Toole and Warren, 1993; MacDonald, 1993; MacDonald and Wunsch, 1996; Ganachaud, 1999). Shriver and Hurlburt (1997; also see Goodman, 1998) find profound effects on the Indonesian ThroughFlow (ITF) if model NADW formation is shut down, implying that even the far-off Atlantic thermohaline budgets may be related to ITF (Gordon, 1986). Increased oceanic heat and freshwater flux into the Indian Ocean at the expense of the Pacific affect atmosphere–ocean coupling with potential impacts on the ENSO and monsoon phenomena. Webster *et al*. (1998) state that the ITF heat flux '… is comparable to the net surface flux over the northern Indian Ocean and a substantial fraction of the heat flux into the western Pacific warm pool … it would appear that the throughflow is an integral part of the heat balances of both the Pacific and Indian Oceans.'

Model research reveals dependence of Pacific and Indian Ocean SST and upper-layer heat storage on the throughflow (Hirst and Godfrey, 1993; Verschell *et al*., 1995; Murtugudde *et al*., 1998). The Indian and Pacific Oceans would be very different if the ITF were zero (MacDonald, 1993). Maes (1998) finds that the effect of a zero ITF would raise sea level in the Pacific and lower sea level in the Indian by 2–10 cm. Schneider (1998) shows that the presence of the throughflow shifts the warmest SST and associated atmospheric convective region towards the west, relative to a no-throughflow condition.

The ITF waters are drawn from the Mindanao (North Pacific thermocline) and Halmahera (South Pacific thermocline) eddies at the Pacific entrance to the Indonesian Seas, between the Philippines and New Guinea. Models show that the ITF source water (North Pacific versus South Pacific) depends upon land geometry and the tropical Pacific wind fields (Nof, 1996; Morey *et al*., 1999; Wajsowicz, 1999). Observations show that the ITF is composed mostly of North Pacific thermocline and intermediate water flowing through Makassar Strait (Fine, 1985; Ffield and Gordon, 1992; Gordon, 1995; Gordon and Fine, 1996). Traces of what may be low-salinity Sulu Sea water are found in the Makassar Strait thermocline in the boreal winter season (Ilahude and Gordon, 1996). Wajsowicz's (1996) model shows that the westernmost deep channel, Makassar Strait, carries the bulk of ITF. While some Makassar throughflow exits the Indonesian Sea within Lombok Channel (Murray and Arief, 1988), most turns eastward within the Flores Sea to enter the Banda Sea before entering the Indian Ocean (Gordon and Fine, 1996). In the deep channels east of Sulawesi, South Pacific water infiltrates (isopycnally) into the lower thermocline of the Banda Sea and dominates the deeper layers through density-driven overflow (Van Aken *et al*., 1988; Gordon and Fine, 1996; Hautala *et al*., 1996; Ilahude and Gordon, 1996).

The Indonesian Seas are not a passive channel linking the two oceans: within the seas the ITF thermal and salinity stratification and the SST are significantly modified by tidal and wind-induced mixing and by sea–air fluxes (Ffield and Gordon, 1992, 1996). The various Pacific water masses composing the ITF are altered, so that the thermohaline profile of ITF entering the Indian Ocean is quite different from that of the source Pacific water masses.

#### *4.7.3.2 Transport*

ITF transport estimates based on observations (**Fig. 4.7.1d**, see Plate XX), models and conjecture range from near zero to 30 Sv (Wyrtki, 1961; Godfrey and Golding, 1981; Murray and Arief, 1988; Godfrey, 1989; Kindle *et al*., 1989; Cresswell *et al*., 1993; Hirst and Godfrey, 1993; MacDonald, 1993; Toole and Warren, 1993; Molcard *et al*., 1994, 1996, in press; Miyama *et al*., 1995; Godfrey, 1996; Gordon *et al*., 1997, 1999a; Potemra *et al*., 1997; Shriver and Hurlburt, 1997; Gordon and McClean, 1999; Gordon and Susanto, 1999; Potemra, 1999).

As part of the US–Indonesian Arlindo programme, velocity and temperature were measured at various depths at two moorings within a 45-km wide constriction of the Makassar Strait near 3°S between December 1996 and June 1998 (Gordon *et al*., 1998, 1999a; Gordon and Susanto, 1999; Ffield *et al*., 2000). The 1997 average southward transport with Makassar Strait is 9.3 Sv, with a range of about $\pm 2.5$ Sv depending on how the surface flow is taken into account (Gordon *et al*., 1999a). Temperature and salinity data obtained during the Arlindo cruises (five cruises from 1991 to 1998) indicate there is no return to the Pacific Ocean of Makassar North Pacific-derived thermocline water along a route east of Sulawesi Island (Gordon and Fine, 1996; Gordon and Fine, in prep; in fact during the La Niña condition of December 1996, a time of strong ITF, see Section 4.7.3.4, additional North Pacific thermocline water may enter the Banda Sea from the north).

Measurements in the Lombok Strait (Murray and Arief, 1988; Murray *et al*., 1989) from January 1985 to January 1986 show an average transport of 1.7 Sv, with a maximum of 4.0 Sv towards the Indian Ocean during July and August, and less than 1 Sv from December 1985 to January 1986. The mean transport between the sea surface and 1250 m in the Timor Passage (between Timor and Australia) measured from March 1992 to April 1993 is 4.3 Sv (Molcard *et al*., 1996). Cresswell *et al*. (1993) using CTD and ADCP sections obtained in October 1987 and March 1988 across Timor Passage find 7 Sv flowing towards the Indian Ocean. Molcard *et al*. (in press) find a range of transport within Ombai Strait (north of Timor, between Timor and Alor Island) during 1996 (December 1995 to December 1996) of 3–6 Sv, depending on the assumed cross-strait shear. While caution is urged as these time series measurements were made at different times and ENSO phases, Makassar Strait transport is comparable to the transport sum of 10.5 Sv through the passages of the Lesser Sunda Island chain.

The various streams of the ITF merge to enter the South Equatorial Current of the Indian Ocean. Within the upper 400 m passing across a section between Java and Australia, the ITF is estimated from 1983 to 1989 XBT data (Meyers *et al*., 1995; Meyers, 1996) to be 5 Sv, with a 12 Sv August–September maximum, and near zero transport in May–June and again in October–November. Quadfasel *et al*. (1996) find that of the 22 Sv transport within the upper 470 m of the South Equatorial Current in the eastern Indian Ocean in October 1987, approximately 9 Sv is derived from the ITF. Gordon *et al*. (1997), using five meridional WOCE sections in the Indian Ocean, find that the Indian Ocean South Equatorial Current transfers on average 9 Sv of Indonesian Throughflow water westward within the Indian Ocean. Fieux *et al*. (1994, 1996) find 18 Sv passing into the Indian Ocean between Java and Australia in August 1989 and 2.6 Sv headed eastward in February 1992. The Indian Ocean South Equatorial Current estimates support an ITF transport of 10 Sv.

Inverse (and related) solutions also suggest mean values for the ITF transport. Piola and Gordon (1984) find 14 Sv, while Toole and Warren (1993) deduce a throughflow of 6.7 Sv. Robbins and Toole (1997), using the same data as Toole and Warren but with a silica constraint, find a ITF transport of 9.9 Sv. MacDonald and Wunsch (1996) arrive at a ITF of 7 Sv with a global inverse solution. de las Heras and Schlitzer (1999) calculate the ITF as 13.2 Sv. Ganachaud (1999) finds a ITF of $15 \pm 3$ Sv.

#### *4.7.3.3 ITF and ENSO*

Observational and model studies suggest the ITF transport sways in tune with ENSO: larger transport during La Niña condition, smaller transport during El Niño (Kindle *et al*., 1989; Bray *et al*., 1996; Fieux *et al*., 1996; Gordon and Fine, 1996; Meyers, 1996; Potemra *et al*., 1997). The high-resolution POP model (Gordon and McClean, 1999) yields a 12 Sv annual average during La Niña and 4 Sv average during El Niño. The Arlindo mooring observations within the Makassar Strait, which span the entire cycle of the strong 1997/1998 El Niño, find a correlation ($r = 0.73$) between Makassar transport and ENSO (Gordon *et al*., 1999a; though the time-series is far too short to say this with assurance). During the El Niño months December 1997 to February 1998 the transport average is 5.1 Sv, while during the La Niña months of December 1996 to February 1997 the average is

12.5 Sv, a 2.5-fold difference. Most of the remaining variance of ITF transport once ENSO effect is removed is explained by the annual cycle, with a June maximum and December minimum (Gordon *et al.*, 1999a), and by intraseasonal events (Sprintall *et al.*, 2000; Susanto *et al.*, 2000).

As 1997 was for the most part an El Niño year, when ITF is expected to be smaller than average, the climatic ITF within the Makassar Strait is expected to be larger than the 1997 mean of 9.3 Sv. Assuming that the relationship of the Makassar ITF to ENSO is defined by the 1.6-year record (which without a longer time series is a crude approximation), a climatic mean for the ITF may be expected to be 13 Sv (from **Fig. 4b** of Gordon *et al.*, 1999a). An ITF transport of 10–15 Sv seems like a fair number to use, at least until observations more accurately determine the ITF mean and variability.

#### 4.7.3.4 ITF transport profile

The Arlindo Makassar Strait measurements suggest a complex vertical profile of transport, with implications for interocean thermohaline fluxes and mass budget of the western tropical Pacific warm pool water: the most persistent and strongest ITF occurs within the thermocline and not within the warm surface layer (Gordon and Susanto, 1999; Gordon *et al.*, 1999a). The data indicate frequent occurrence of maximum southward speeds within the mid to lower thermocline. The subsurface maximum occurs during times of large transport, from April to September 1997 and again in April 1998 to the end of the record in June 1998.

An upward-looking ADCP at 150 m on the Arlindo Makassar Strait moorings provided a record of surface layer flow from 1 December 1996 to 1 March 1997. This time series shows increasing southward speeds with increasing depth, with surface flow varying from near zero to northward. This was not a local wind effect, as the NSCAT-measured winds during this period were directed towards the south. Model results suggest similar temporal dependence. Masumoto and Yamagata (1993) with the GFDL (Geophysical Fluid Dynamics Laboratory) model forced by Hellerman and Rosenstein (1983) winds show northward surface flow within Makassar during the winter months. Using a model driven by ECMWF (European Centre for Medium Range Weather Forecasting) winds, Potemra *et al.* (1997) find that Ekman transport in the Indonesian Seas is directed towards the Pacific in the winter months, with strong Indian Ocean-bound Ekman transport in summer. The 1/6° resolution POP (Parallel Ocean Program) model display a surface (upper 100 m) flow towards the north in Makassar Strait during the winter, and towards the south in the summer months (McClean, personal communication, February 1999).

The reality and causes of seasonal oscillations of the surface flow must be further investigated. Possible candidates are: strong monsoonal zonal wind across the southern boundary of the Makassar Strait (Java and Flores Seas) relative to the weak monsoonal winds in the northern boundary of the Makassar Strait (Sulawesi Sea); Ekman pumping along the southern coast of the Sunda Islands (Potemra, 1999); or a buoyancy effect induced by the enormous amount of very-low-salinity Java Sea water injected into the Flores Sea and southern boundary of the Makassar Strait during the boreal winter.

#### 4.7.3.5 Thermohaline fluxes

The Arlindo Makassar Strait time series reveals that the ITF transport is linked to thermocline depth: transport is smaller and thermocline shallower during El Niño (Bray *et al.*, 1996, 1997; Meyers, 1996; Ffield *et al.*, 2000). The correlation between variability in the average thermocline temperature to variability in the southward Makassar transport is $r = 0.67$ (Ffield *et al.*, 2000). Using nearly 15 years of XBT data, Ffield *et al.* (2000) show that the Makassar upper thermocline temperature is highly correlated with ENSO: 0.77 for the Southern Oscillation Index; −0.80 for NINO3 SST anomaly, and −0.82 for NINO4 SST anomaly. The correlations increase when the ENSO time series are lagged a month or so. Thus the Makassar temperature field – when coupled with the throughflow – transmits equatorial Pacific El Niño and La Niña temperature fluctuations into the Indian Ocean.

An estimate of the internal energy transport (Warren, 1999) for the Makassar Strait is made by Ffield *et al.* (2000) by integrating the product of temperature, volume transport, density and specific heat in the upper 400 db of the water column. The 0 to 150 db temperature is obtained by subtracting the depth-weighted 150 to 400 db temperature time series from the full-depth temperature time series estimated from the travel time measured by Inverted Echo Sounders. The 1997 Makassar Strait internal energy transport is 0.50 PW (petawatts). If the Makassar temperatures are referenced to 3.72°C, as in Schiller *et al*. (1998), the 1997 internal energy transport is reduced to 0.39 PW. The ENSO influence on the internal energy transport may be estimated: 0.63 PW during the La Niña months of December 1996 through February 1997, and 0.39 PW during the El Niño months of December 1997 through February 1998.

The average transport-weighted temperature determined by the Arlindo moorings is approximately 11.5°C with an average salinity of 34.45 (K. Vranes, Columbia graduate student). This is cooler and saltier than the ITF characteristics envisioned by Piola and Gordon (1984; ITF of 33.6) and Toole and Warren (1993; ITF temperature of 24°C). Toole and Warren (1993) deduce a throughflow of 6.7 Sv with salinity of 34.5. They assign it with a temperature of 24°C, which yields a heat-flux divergence between Indonesia and 32°S of 0.98 PW. The heat and freshwater divergence within the Indian Ocean north of 32°S depends on the temperature and salinity of the net poleward flow across 32°S. Using the Toole and Raymer (1985) Fig. 3 and their mean temperature and salinity along 32°S across the Indian Ocean of 5.71°C and 34.80, together with the ITF characteristics found by the Arlindo moorings, yields an ITF heat flux of roughly 0.27 PW and a net precipitation minus evaporation plus runoff over the Indian Ocean (north of 32°S) of nearly −0.3 Sv. As the ITF water is not likely to be cooled to 5.71°C, this value represents a maximum heat flux. An estimate of minimum heat flux may be calculated by assuming the ITF water eventually passing poleward within the Agulhas Current across 32°S confined to roughly the same density interval as the ITF in the Indonesian passages. This yields a value statistically equal to zero. This means that the ITF heat may be eventually lost south of 32°S, consistent with the model results of Hirst and Godfrey (1993) and of Allan *et al*. (1995), who find much enhanced ocean-to-atmosphere heat flux along 40°S when the ITF is included.

## 4.7.4 The Agulhas Retroflection

### *4.7.4.1 Introduction*

A third interocean exchange route considered as important to larger-scale thermohaline circulation lies not within the confines of a channel, but rather in the gap between the southern shores of Africa and the ACC, a gap occupied by the Agulhas Retroflection (see Lutjeharms, 1996, **Fig. 1d**). The Agulhas Current flows westward along the southern rim of Africa, with a transport approaching 100 Sv. Rather than continuing into the South Atlantic, Agulhas water curls back to the Indian Ocean, feeding the eastward-flowing Agulhas Return Current near 40°S, the ACC main axis falling further south, near 48°S (Read and Pollard, 1993). However, not all of the Agulhas water turns back to the Indian Ocean. That part that does not turn back passes into the South Atlantic, in what is often called Agulhas leakage. Of all of the interocean exchanges, the Agulhas leakage into the Atlantic has drawn the most attention (and controversy) in terms of its role in NADW formation (Gordon, 1996a).

Recent WOCE-generated advances in our understanding of the Indian–South Atlantic exchanges are reported in two recent collections: Wefer *et al*. (1996) and a special section of the *Journal of Geophysical Research* (Gordon *et al*., 1999b). The articles by De Ruijter *et al*. (1999), Witter and Gordon (1999), Garzoli *et al*. (1999), Arhan *et al*. (1999) and McDonagh and Heywood (1999) are particularly relevant to the topic of invasion of Indian Ocean water into the Atlantic. Also see Rintoul *et al*., Chapter 4.6 and the model results of Semtner and Chervin (1992), Boddem and Schlitzer (1995), Cai and Greatbatch (1995), Döös (1995), Lutjeharms and Webb (1995) and Florenchie and Verron (1998). How models 'handle' the Agulhas leakage depends very much on their eddy-resolving characteristics (Marsh *et al*., 2000). The De Ruijter *et al*. (1999) paper provides an excellent review of interocean exchange afforded by the Agulhas Retroflection.

### *4.7.4.2 Transport*

*Benguela Current*

Gordon *et al*. (1992), using oxygen and CFC data collected during the South Atlantic Ventilation Experiment cruise no. 4 (SAVE-4) from 7 December 1988 to 15 January 1989 (not 1989 and 1990, as stated in Gordon *et al*., 1992), find that of the 25 Sv Benguela Current geostrophic transport, 15 Sv are drawn from the Indian Ocean. Nine Sv (2 Sv of thermocline water warmer than 9°C and 7 Sv of lower thermocline and Antarctic Intermediate

Water) of this Indian Ocean water pass into the North Atlantic within the North Brazil Coastal Current. The rest of the Indian Ocean water entering the South Atlantic follows the subtropical gyre, eventually to return along the horizontal plane to the Indian Ocean along a path just south of the Agulhas Return Current. As the Indian Ocean thermocline and intermediate layers are saltier than the other major provider of the Benguela Current water (the South Atlantic Current defining the polar limb of the subtropical gyre), the Agulhas leakage also adds salt to the South Atlantic. The leakage is also warmer than the waters of the South Atlantic Current, and hence also adds heat to the Atlantic Ocean.

Saunders and King (1995), using WHP section A11 (roughly along 40°S, ending at 30°S at Africa), find a northward transport of 10 Sv of thermocline water ($<\sigma_0$ of 26.8) drawn from the Agulhas Retroflection and 5 Sv of AAIW. Garzoli *et al.* (1996) estimate the Benguela Current transport, based on a 16-month (1992–93) time series of Pressure and Inverted Echo Sounder (PIES) and current meter data at 30°S, as 16 Sv. Applying the property-based estimate of the percentage of Indian Ocean water found by Gordon *et al.* (1992), the Indian Ocean influx would be on average 10 Sv, 6 Sv of which enter into the upper limb of the Atlantic meridional overturning circulation. During the mooring deployment discussed by Garzoli *et al.* (1996), three oceanographic sections were obtained across the Benguela Current near 30°S. While the sample number is small, higher transport is associated with a more saline thermocline and intermediate layer, suggesting that variable Benguela Current transport is due to a variable Agulhas leakage.

Holfort and Siedler (2000) in an inverse solution find 9 Sv of warm surface water (Agulhas leakage) and 6 Sv of cold AAIW pass northward across 30°S in the South Atlantic. This is close to the values found by Saunders and King (1995). Their partitioning between warm and cold route contrasts with that of Schlitzer (1996) of 2 Sv of Agulhas leakage to 11.9 Sv of AAIW. Stutzer and Krauss (1998) assimilate drifter trajectories into a model for the South Atlantic. They find that the Benguela Current, with its Indian Ocean component, directly feeds the South Equatorial Current, with bifurcation near 10°S along the Brazilian coast. Clearly some Indian Ocean water is deflected into the North Brazil Current and northern hemisphere.

*Agulhas eddies*

The primary means by which the Agulhas injects Indian Ocean water into the Benguela Current occurs within large anticyclonic eddies. These energetic eddies of Agulhas (Indian Ocean) water can be traced across the South Atlantic Ocean (Byrne *et al.*, 1995). Interannual variability of the width and latitude of the transocean corridor occurs, with expansion and contraction of the South Atlantic subtropical gyre (Witter and Gordon, 1999). Effects of seafloor topography on the eddy propagation is clearly evident in observations and models (Byrne *et al.*, 1995; Kamenkovich *et al.*, 1996; Florenchie and Verron, 1998). On average six Agulhas eddies enter the South Atlantic each year, injecting from 3 to 9 Sv into the South Atlantic Ocean. In transit they dissipate, losing their core of Indian Ocean water (Byrne *et al.*, 1995) into the Benguela Current and South Equatorial Current.

#### 4.7.4.3 Heat and freshwater fluxes

The South Atlantic contributes to the overall high thermocline salinity of the Atlantic Ocean in two ways: the ACC exports fresh water from the South Atlantic subtropics (Gordon and Piola, 1983); warm saline Indian Ocean thermocline water is injected into the Benguela Current (Gordon *et al.*, 1992). Agulhas leakage occurs in the form of Agulhas rings shed from the Agulhas Current's retroflection and surface water filaments (Lutjeharms, 1996). In addition, branches of the Agulhas Current may flow directly into the Benguela Current between the Africa and the main offshore corridor that carries the eddies towards the northwest (Gordon *et al.*, 1992; Garzoli *et al.*, 1996). These branches of Indian Ocean water lost from the core of the eddies are carried into the South Equatorial Current of the South Atlantic, part of which passes into the northern hemisphere.

Byrne *et al.* (1995) estimate that the average Agulhas eddy sampled within the southeastern South Atlantic transfers at least 0.8 Sv of Indian Ocean water into the South Atlantic. Two newly formed Agulhas eddies found near the retroflection each contributed 1.4 and 1.8 Sv, respectively, of Indian Ocean water (Byrne *et al.*, 1995). Clement and Gordon (1995) find that three Agulhas eddies found within the Cape Basin (east of Walvis Ridge) each transferred between 0.45 and 0.90 Sv of Indian Ocean thermocline and intermediate water. As there are on average six such eddies per year, the total flux of Indian Ocean water above 1000 m introduced into the Atlantic by the eddy mechanism may be from 5 to 10 Sv. Van Ballegooyen *et al.* (1994) finds that Agulhas eddies contribute salt and heat to the South Atlantic thermocline waters at a rate of up to $2.5 \times 10^6\,kg\,s^{-1}$ (salt) and 0.045 PW (heat), assuming a yearly total of six eddies. Agulhas filaments may provide only 13% of the total interocean salt flux, the heat being lost quickly to the atmosphere (Lutjeharms and Cooper, 1996). Garzoli *et al.* (1996) and Duncombe Rae *et al.* (1996) find that the eddies measured during the BEST experiment (1992–93) contributed 0.007 PW, $4.5–6.7 \times 10^5\,kg\,s^{-1}$ and 2.6–3.8 Sv of heat, salt and volume flux, respectively, into the South Atlantic Ocean.

Garzoli *et al.* (1999) concluded from a detailed survey of three Agulhas eddies that a reasonable estimate for mass and heat flux from the Indian Ocean to the Atlantic is 6–10 Sv and 0.006 PW. The decrease of Indian Ocean core water observed as an Agulhas eddy dissipates indicate a slow transfer of Indian Ocean water from the confines of the eddy to the Benguela Current. In this way Indian Ocean water injected into the South Atlantic within an eddy may not accompany the dynamic feature of the eddy across the South Atlantic near 30°S (Byrne *et al.*, 1995; Witter and Gordon, 1999), but rather may be advected to the northwest within the general circulation of the Benguela Current and South Equatorial Current, as shown schematically by Gordon *et al.* (1992). The interaction of an Agulhas eddy with the atmosphere, seafloor topography (Kamenkovich *et al.*, 1996) and surrounding South Atlantic water strongly control its trajectory and modification (Arhan *et al.*, 1999).

Weijer *et al.* (1999) find that heat and salt flux associated with South Atlantic interocean exchanges, and in particular the ratio of thermocline to intermediate components (lateral boundary buoyancy profile), has a strong influence on the Atlantic meridional overturning circulation. An important aspect of the injection of Indian Ocean salt into the South Atlantic at the Agulhas Retroflection may be related to the '…shape of the salt flux profile…', but the more important impact of Agulhas leakage to Atlantic meridional overturning circulation may be due to the Indian Ocean heat flux, which will drive further evaporation in the South Atlantic.

There is ambiguity in the heat flux calculations associated with Agulhas leakage, as authors use different reference temperatures. To determine interocean heat flux one needs to identify the temperature of the water that leaves the Atlantic Ocean, balancing the mass injection of the Agulhas leakage. There are two reasonable options (Gordon, 1985): on the horizontal plane, with the transfer of cooler South Atlantic Current upper-layer water into the Indian Ocean south of the Agulhas Return Current; or on the vertical plane, by export of (2°C) NADW into the Indian and Pacific Oceans. Transport-weighted temperature (relative to 1500 m) was estimated from the SAVE-4 section across the South Atlantic Current crossing 10°W (used by Gordon *et al.*, 1992) as 10°C. The transport-weighted Agulhas transport (relative to 1500 m) taken between the station pair 49 and 50 in Gordon *et al.*, (1987) is calculated as 15°C. The heat flux from the Indian Ocean to the Atlantic for an Agulhas leakage of 15 Sv (upper 1500 m, Gordon *et al.*, 1992) for closure within the upper 1500 m is 0.3 PW (this is a larger value than calculated by Gordon (1985), where the smaller temperatures difference from opposing sides of a detached Agulhas eddy was used). For closure by the colder NADW the heat flux is 0.8 PW. The corresponding values for 5 Sv and 10 Sv Agulhas leakage, respectively are 0.1, 0.3 PW and 0.2, 0.6 PW. Estimates of the northward heat flux across the South Atlantic subtropics (18°S to 30°S) ranges from 0.1 to 0.9 PW, with error estimated in excess of 0.2 PW (MacDonald and Wunsch, 1996; Ganachaud, 1999; Holfort and Siedler, 2000). Heat flux values of less than 0.3 PW indicate that Agulhas leakage is predominately balanced by export of Atlantic water within the upper 1500 m, while heat flux values greater than 0.3 PW suggest significant involvement with the Atlantic meridional overturning circulation. The two methods for closure are not mutually exclusive and their relative importance may be time variable.

Within 50 km of the coast the Agulhas Current intermediate water, in the 4–8°C interval, is more saline than at points further offshore (Beal and Bryden, 1999). Remnants of Red Sea water flow southward are observed hugging the continental slope to at least the southern coast of Africa (Gordon, 1986; Beal *et al.*, 2000), whereas offshore of the Agulhas axis, within the anticyclonic zone, the intermediate depths are dominated by lower-salinity AAIW coming from the subtropical

Indian Ocean. Beal *et al*. (2000) conclude that all of the Red Sea overflow into the Indian Ocean is eventually exported by the western boundary current. As Red Sea water is not observed within the Agulhas Return Current, either its signature is removed by mixing or it fully contributes to the Agulhas leakage. Possibly the inshore component of the Agulhas preferentially contributes to the interocean exchange, and to the Atlantic's high salinity.

### 4.7.5 Discussion

Interocean exchanges have been subject to much attention in recent years (Schmitz, 1995, 1996a,b). The ACC, Bering Strait, Indonesian Seas and Agulhas Retroflection (and perhaps the westward flow immediately south of Australia) offer pathways for interocean exchange. Interocean fluxes of mass, heat and fresh water are expected to vary across the full range of temporal scales in concert with climate variability. As they vary, heat and freshwater budgets of neighbouring oceans change; temperature and salinity anomalies from long-term means can develop. These anomalies may play a role in the climate phenomena (oscillations) associated with each ocean basin. Peterson and White (1998) find that heat and freshwater anomalies forming in the western subtropical South Pacific Ocean spread eastward within the ACC entering the subtropical gyres of the Southern Hemisphere. The migration of these anomalies is linked to the 3- to 5-year Antarctic Circumpolar Wave (White and Peterson, 1996).

Transport of Pacific waters entering the Indian Ocean via the Indonesian Seas displays strong dependence of the phase of ENSO. During El Niño the Indonesian Throughflow is greatly reduced, delivering less heat and fresh water into the Indian Ocean. Might these variations introduce heat and freshwater anomalies into the Indian Ocean thermocline, linking ENSO and monsoon climate phenomena? The transfer of subtropical Indian Ocean water into the South Atlantic also displays variability. The transport of the Benguela Current may be modulated by Agulhas leakage (Garzoli *et al*., 1996). The pathway of Agulhas eddies across the South Atlantic varies interannually, along with pulsations in the form of the subtropical gyre (Witter and Gordon, 1999), influencing the distribution of heat and fresh water within the South Atlantic, and perhaps altering the access of the Agulhas leakage to the pathways along the Brazilian coast leading into the northern hemisphere (Stramma and Schott, 1996; Schott *et al*., 1998).

Roach *et al*. (1995) find that there can be interannual variations of up to 1 ppt in salinity of the Bering Strait throughflow. Niebauer (1998) discusses the variability of wind over the Bering Sea to ENSO and how this relationship changed in the 'regime shift' of the later 1970s. Changing wind fields and sea ice distribution in the Bering Strait region may have a downstream effect on the Arctic freshwater budget and the flux of fresh water through Fram Strait, affecting NADW formation within the Greenland, Norwegian and Labrador Seas. As the Bering Strait did not exist during the lowered sea level of the glacial epochs, a vital freshwater source for the Arctic was severed. While other parts of the Arctic hydrological cycle would be greatly altered by the glacial condition, reduced inflow of Pacific fresh water into the Arctic pycnocline may be expected to reduce its stability and allow greater vertical heat flux from the warmer deep water to the sea ice cover.

The Atlantic Ocean receives low-salinity water from the Arctic and saline water from the Indian Ocean in association with Agulhas leakage (Gordon *et al*., 1992; De Ruijter *et al*., 1999). The former acts to attenuate convection in the northern North Atlantic (Zaucker *et al*., 1994), while the latter has the opposite effect. On entering the South Atlantic a density anomaly of Indian Ocean water is only small, due to the counteracting effects of the heat and salt anomalies on the density. However, the heat is quickly lost to the atmosphere while the salt remains in the water column (Gordon *et al*., 1992). According to modelling studies the effect of the warm and salty Indian Ocean source is to strengthen and stabilize the northern meridional overturning of the Atlantic (Weijer *et al*., 1999; Weijer, 2000; Weijer *et al*., 2000a,b), while the effect of the Bering Strait (Arctic source) freshwater flux is to weaken the northern overturning. In the present-day climate state the Indian Ocean source effect dominates over the northern freshwater fluxes. Shutting down the source of Indian Ocean water brings the ocean circulation close to a state where it can switch between two equilibrium conditions: one with the northern-driven overturning, the other with southern-driven overturning (see Rahmstorf, 1995, 1996). With the Agulhas leakage reduced to near zero, the Bering Strait takes on greater significance, with relatively small variations in Bering Strait inflow (or other freshwater sources in the north) initiating switches from one state to the other, perhaps triggering large climate fluctuations.

It is clear that distribution of heat and fresh water associated with NADW formation has a major impact on North Atlantic (and points beyond?) climate. The process is very robust, as even massive introductions of fresh water do not permanently shut down the North Atlantic thermohaline overturning circulation: it's a salty ocean and it wants to stay that way. Why? Is it locally maintained (certainly there is a lot of positive feedback), or might there be a global factor such as the pattern of interocean exchange (export of fresh water from the Atlantic by the ACC and by introduction of Indian Ocean salt through Agulhas leakage)? NADW spreads into the ACC, entering the southern ocean thermohaline overturning cell, eventually contributing to AAIW, spreading into each ocean. Most of the AAIW upwells into the thermocline of the major oceans. Its NADW-derived component must weave its way back to the North Atlantic. Much of AAIW directly enters the South Atlantic upwells, mostly in the tropics (You, 1999), to enter the North Atlantic thermocline. Some AAIW no doubt upwells in the Pacific and Indian Oceans, to be converted to thermocline water, which then passes, with increasing salinity, through the Indonesian Seas and around the rim of southern Africa. Strong transients overshadowed by other elements of the regional circulation complicate the pathway, but on the decadal and longer scales, its impact on climate is felt.

Climate models, in their quest to simulate the variability, must not only get mass and thermohaline fluxes correct within the major ocean basins, but they must also simulate the fluxes in the constricted pathways connecting these oceans (see, for example, Ribbe and Tomczak, 1997). This is a very challenging task because of the complex boundary conditions within the passages, temporal variability and lack the observational base to enable evaluation of the model results.

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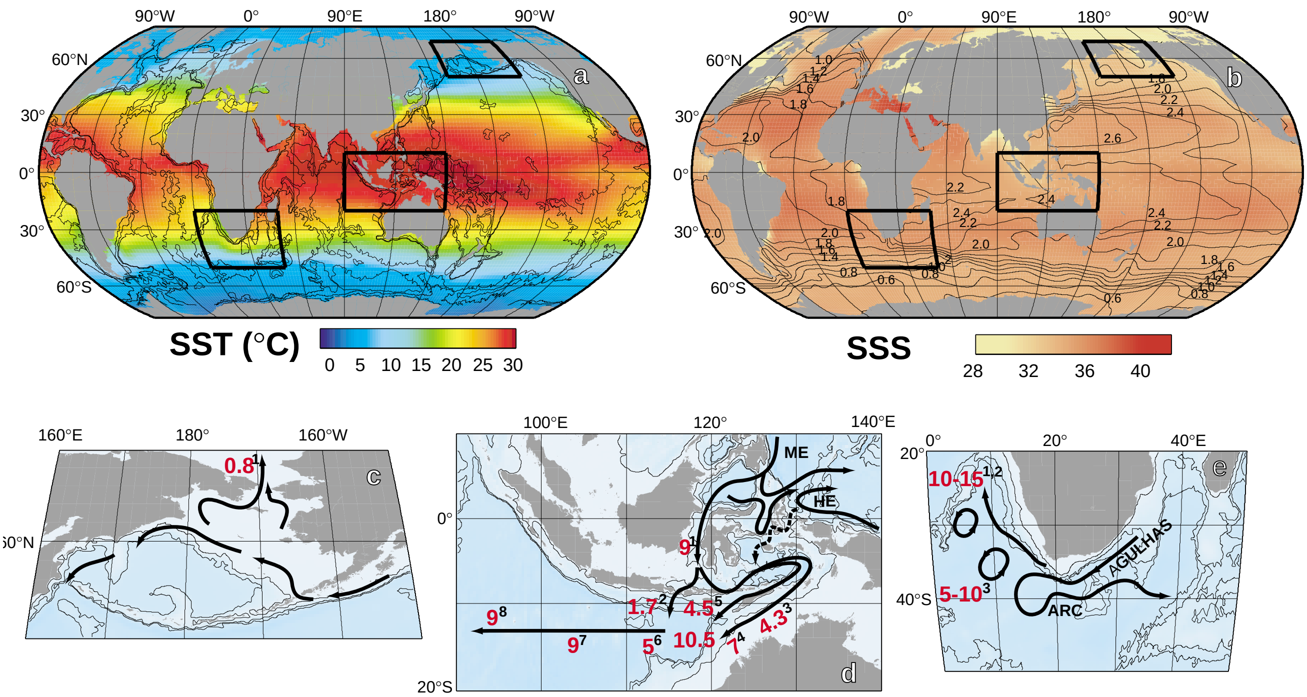


**Plate 4.7.1**

**(a)** Sea Surface Temperature (SST) and seafloor topography (500, 3000, 4000 m isobaths are shown). SST data are derived from Conkright *et al*. (1998).

**(b)** Sea Surface Salinity (SSS) and dynamic height anomaly (dynamical metres) of the sea surface relative to 2000 decibars. SSS and dynamic height anomaly are derived from Conkright *et al*. (1998).

**(c)** Bering Strait: transports in $10^6$ $m^3s^{-1}$ are given in red. Superscript refers to reference source: 1: Roach *et al*. (1995).

**(d)** Indonesian Seas: transports in $10^6$ $m^3s^{-1}$ are given in red. The 10.5 Sv is the sum of the flows through the Lesser Sunda passages. ME Mindanao Eddy; HE Halmerhera Eddy. Superscripts refer to reference sources: 1, Makassar Strait transport in 1997 (Gordon *et al*., 1999a); 2, Lombok Strait (Murray and Arief, 1988; Murray *et al*., 1989) from January 1985 to January 1986; 3, Timor Passage (between Timor and Australia) measured in March 1992 to April 1993 (Molcard et al., 1996); 4, Timor Passage, October 1987 and March 1988 (Cresswell *et al*., 1993); 5, Ombai Strait (north of Timor, between Timor and Alor Island) from December 1995 to December 1996 (Molcard *et al*., 2000); 6, between Java and Australia from 1983 to 1989 XBT data (Meyers *et al*., 1995; Meyers, 1996); 7, upper 470 m of the South Equatorial Current in the eastern Indian Ocean in October 1987 (Quadfasel *et al*., 1996); 8, average ITF within the South Equatorial Current defined by five WOCE WHP sections (Gordon *et al*., 1997). Not shown is the overflow of dense Pacific water across the 1940-m deep Lifamatola Passage (1°49'S; 126°57'E) into the deep Banda Sea, which may amount to about 1 Sv (Van Aken *et al*, 1988).

**(e)** Agulhas Retroflection: transports in $10^6$ $m^3s^{-1}$ are given in red. ARC, Agulhas Return Current. Superscripts refer to reference sources: 1, Thermocline water (<$\sigma_0$ of 26.8) passing northward across WHP section A11 (1993) (Saunders and King, 1995); Holfort and Siedler, 2000); 2, Benguela Current, January 1989 (Gordon *et al*., 1992); 3: Agulhas Eddies (Byrne *et al*., 1995).

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