The deep meridional overturning circulation in the Indian Ocean inferred from the GECCO synthesis

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The deep time-varying meridional overturning circulation (MOC) in the Indian Ocean in the German “Estimating the Circulation and Climate of the Ocean” consortium efforts (GECCO) ocean synthesis is being investigated. An analysis of the integrated circulation suggests that, on time average, 2.1 Sv enter the Indian Ocean in the bottom layer (>3200 m) from the south and that 12.3 Sv leave the Indian Ocean in the upper and intermediate layers (<1500 m), composed of the up-welled bottom layer inflow water, augmented by 9.6 Sv Indonesian Throughflow (ITF) water. The GECCO time-mean results differ substantially from those obtained by inverse box models, which being based on individual hydrographic sections and due to the strong seasonal cycle are susceptible to aliasing.

The GECCO solution shows a large seasonal variation in its deep MOC caused by the seasonal reversal of monsoon-related wind stress forcing. The associated seasonal variations of the deep MOC range from −7 Sv in boreal winter to 3 Sv in summer. In addition, the upper and bottom transports across the 34°S section show pronounced interannual variability with roughly biennial variations superimposed by strong anomalies during each La Niña phase as well as the ITF, which mainly affect the upper layer transports. On decadal and longer timescale, the meridional overturning variability as well as long-term trends differs before and after 1980. GECCO shows a stable trend for the period 1960–1979 and substantial changes in the upper and bottom layer for the period 1980–2001.

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1. Introduction

A deep meridional inflow is one of the characteristic features at the southern boundary of the Indian Ocean, where North Atlantic Deep water (NADW), Antarctic Bottom Water and Circumpolar Deep Water enter the Indian Ocean in the western basin off Madagascar and off the coast of East Africa, and in the east along the Ninety East Ridge (Mantyla and Reid, 1995; Schott and McCreary, 2001, hereafter denoted as SMCO1). This water forms the lower limb of the deep MOC in the Indian Ocean, and mixes and upwells to the intermediate and upper layers before it exits the Indian Ocean again across the same southern boundary. The pathway of the deep water in the Indian Ocean is fundamentally restricted by numerous topographic ridges as illustrated in Fig. 1. Despite much progress since the 1980s, the description of the time-mean deep MOC and its temporal variability are still a matter of debate.

Early studies of the deep Indian Ocean MOC by Toole and Warren, 1993 and its later re-evaluation by Robbins and Toole (1997) provided an estimate of 12 Sv below 2000 m based on geostrophic transports across individual transoceanic hydrographic sections along 32°S. Later results from inverse box

By means of an extended EOF analysis, the importance of Ekman dynamics as driving forces of the deep MOC of the Indian Ocean on the interannual timescale is highlighted. The leading modes of the zonal and meridional wind stress favour a basin-wide meridional overturning mode via Ekman upwelling or downwelling mostly in the central and eastern Indian Ocean. Moreover, tropical zonal wind stress along the equator and alongshore wind stress off the Sumatra-Java coast contribute to the evolution of the Indian Ocean dipole (IOD) events.
models revealed a similar vertical structure of the overturning transport circulation. However, results differ drastically with respect to the strength of lower limb of the deep overturning circulation below 2000 m (ranging from 5 to 23 Sv) (Macdonald, 1998; Ganachaud et al., 2000; Sloyan and Rintoul, 2001; Ganachaud, 2003; Lumpkin and Speer, 2007; McDonagh et al., 2008). The use of different hydrographic sections and their treatment of diapycnal fluxes in the inverse box model are possible reasons [SMC01].

Ocean general circulation models provide an alternative way to estimate the deep MOC of the Indian Ocean. As an example, Gartemnicht and Schott (1997) provide estimates of the annual–mean meridional stream functions from model results. The resulting maximum inflow amounts to about 3 Sv below 4000 m, which outflows between 800 and 4000 m. Ocean state estimation, based on an adjoint assimilation model constrained by large-scale ocean data sets, provide a time mean strength of 4 Sv for the deep MOC during the period 1992–1997 (Stammer et al., 2002) and an estimate of the inflow below 3000 m of less than 2 Sv (Lee and Marotzke, 1998). In comparison to estimates from observational studies, all model-based results (constrained or unconstrained) are much weaker, the only exception being the regional estimate of 17 Sv obtained by Ferron and Marotzke (2003) by essentially assimilating the same hydrographic sections with the same 4D variational method used by Stammer et al. (2002).

Many modelling efforts investigated the time–mean deep meridional overturning. However, only few studies focused on the time-varying circulation. The stream function of the deep MOC of the Indian Ocean from model results reveals large temporal reversals between monsoon phases (Gartemnicht and Schott, 1997; Lee and Marotzke, 1998), indicating a substantial deep-reaching effect of the seasonally varying monsoon winds on the deep meridional circulation cell [SMC01]. However, not much is known about those variations of the deep MOC on interannual and longer time-scale.

In this study we will use the 50-year long GECCO ocean state estimation results, covering the period from 1952 to 2001, to revisit the estimate of the time-mean deep MOC and to investigate its time-varying features. In particular we will revisit the discrepancies between the model and observations based MOC estimates in view of the GECCO global ocean assimilation product.

The overall structure of this paper is as follows. Section 2 summarizes the GECCO synthesis. Section 3 considers the time-mean deep meridional overturning, including integrated circulations of the upper, intermediate, deep, and bottom layer in the GECCO synthesis. Section 4 discusses the time-varying features of deep MOC in the GECCO synthesis covering from seasonal, interannual to decadal and longer timescales. Section 5 provides a discussion and concluding remarks.

2. Data and model

The ECCO ocean state estimate used in this study is described in detail by Stammer et al. (2004), Köhl et al. (2006) and Köhl et al. (2007). The estimate is available for a quasi-global (±80° latitude) domain with 1° spatial resolution and spans the period 1952–2001. It is obtained by combining most of the global observations available during the entire estimation period with the ECCO model by changing the initial temperature and salinity fields over the full water column and by adjusting the NCEP time-varying surface forcing over the full estimation period so as to simulate best the observed ocean state. In this approach, the adjoint model provides the means to transform the observed ocean data into improvements of the control parameters, which then by driving an ocean model determine the synthesis. The strength of the synthesis is that all data may contribute via the improved parameters to all aspects of the ocean circulation even in areas that are not directly observed. However, a synthesis based on the adjoint method remains essentially an, albeit, improved ocean simulation and shares with a plain forward simulations most of their strengths (conservation principles) and their weaknesses (model biases). The prior forcing fields consists of twice–daily wind stress and daily heat and freshwater flux fields from the National Centre for Environmental Prediction. These forcing fields are adjusted every 10 days by the adjoint method to yield a model state that is dynamically consistent, within given error limits, with the model physics and the assimilated data alike. Details of the 50-year long GECCO optimization are described by Köhl and Stammer (2008a,b). Error estimates of the mean transport values are calculated from the standard deviation of the monthly mean values.

During the assimilation effort, special attention was paid to the mean values of the transports through key passages that are marginally resolved such as Denmark Strait, Florida Strait or the
Indonesian Throughflow (ITF). Resulting transports remain mostly within 3% of the constrained reference run, indicating that these values are not sensitive to the data and are mainly determined by the configuration of the model. We also note that the adjustments of the initial conditions lead to a density field that in some regions is less in equilibrium with the wind forcing than the reference run. The timescales of associated adjustment processes are typically 5–7 years and affect mostly the first 10 years, which were discarded in our analysis, which thus only spans the 42-year period 1960 through 2001. Some adjustment effects on the second decade remain, which is particularly true for the ACC transport for which a longer timescale of more than 10 years applies.

GECCO results are analyzed here with respect to the time-varying volume transports in the Indian Ocean in the same density classes used by Ganachaud and Wunsch (2000) and Ganachaud et al. (2000). Following their work, estimates were obtained in four neutral density classes (Jackett and Mcdougall, 1997) separated by the surfaces $\gamma_n = 27.05$, $\gamma_n = 27.72$ and $\gamma_n = 28.11$ kg m$^{-3}$, which separate water masses in the upper ($\gamma_n < 27.05$), intermediate ($27.05 < \gamma_n < 27.72$), deep ($27.72 < \gamma_n < 28.11$) and bottom layers ($\gamma_n > 28.11$), respectively. Their depth distribution is shown in Fig. 2. Generally, the depth of neutral surfaces of 27.05 and 27.72 are uniform in the region roughly south of 20°S, where they have similar structures with maximum depth south of Madagascar. The depth of both surfaces is shallower than 550 m and 1350 m, respectively, north of 20°S. The depth of the deep layer, 28.11, is larger than 3400 m in the Arabian Basin, around 3350 m in the Somali Basin and 3200–3300 m in the Central Indian Basin and the West Australian Basin.

3. Time-mean deep MOC of the Indian Ocean

The deep overturning in the Indian Ocean is illustrated in Fig. 3 in terms of the time-mean meridional transport stream function. The figure reveals that in the GECCO solution a weak, but noticeable, deep MOC exists which is characterized by a double core in the southern Indian Ocean at depths of 3200 m and 1500 m, respectively. One core is centred on the neutral density surface of 27.72, the other on 28.11. With both cores together, 2.1 Sv enter the Indian Ocean in the bottom layer, upwell in the interior of the Indian Ocean, and exit the Indian Ocean again in the intermediate layer (roughly between 700 m and 1500 m). Estimates of the upper layer seem to be dominated by the ITF transport. Inverse Box models previously resulted in estimates of an overturning transport stream function at 32°S with a similar vertical structure but with a greater magnitude of the deep overturning (Fig. 4), varying from 11.5 Sv within neutral density classes 28.0–28.1 (Ganachaud et al., 2000) to 23 Sv within 28.0–28.3 (Sloyan and Rintoul, 2001). Differences are likely to originate from assumptions and configurations of inverse box models and numerical general ocean circulations models, which we will discuss below in more detail. In contrast, GECCO shows a weak time-mean overturning transport of slightly more than 2 Sv (below 3000 m).

The integrated GECCO time-mean circulation in the upper layer (Fig. 5a) captures all main characteristics of the near-surface flow field of the Indian Ocean as it is described in by Schott et al. (2009), including the ITF, the South Equatorial Current (SEC), the South Equatorial Counter Current (SECC), the Mozambique Current, and the subtropical gyre. The ITF with 9.2 Sv flows westward (Fig. 1) and is in close agreement with the earlier estimate from measurements made prior to 2000 (Gordon et al., 2003) of about 10 Sv, but smaller than the recent estimate of 13.6 Sv from 14 ocean data assimilation products for the 1993–2001 period (Lee et al., 2010) and 15 Sv for the 2004–2006 INSTANT Programme period (Gordon et al., 2010). Part of the ITF joins the SEC, while the rest flows southward to join the Mozambique Current through the Mozambique Channel and along the east coast of Madagascar via the Southeast Madagascar Current (Song et al., 2004; Valsala and Ikeda, 2007). The broad westward SEC brings about 41.5 Sv from the eastern basin to the west and eventually reaches the east coast of Madagascar. At about 17°S, the SEC splits into the Northeast and Southeast Madagascar Current (NEMC and SEMC), which is estimated as 27.4 Sv and 14.1 Sv, respectively. The SECC brings about 11.8 Sv back from the west basin across 80°E to the east, part of which joins the ITF close to Sumatra. As a branch of the NEMC, the Mozambique Current carries 9.9 Sv southward through the Mozambique Channel and feeds into the western boundary current.

The integrated circulation of the intermediate layer mirrors that of the upper layer, albeit with a much weaker strength south of 10°S and without the ITF and the SECC superimposed (Fig. 5b).
Fig. 2. GECCO time-mean (over the period 1960–2001) depth of neutral density surfaces (a) 27.05, (b) 27.72, and (c) 28.11 (\(\gamma^\alpha\) in kg m\(^{-3}\)) in the Indian Ocean.

Only 2.8 Sv contribute to the SEMC and 8.9 Sv across the 60°E westward originating mainly from the west coast of Australia; 3.4 Sv flow southward through the Mozambique Channel. This results in a net 18.4 Sv outflow of the Agulhas Current in the intermediate layer. Combined with the transports in the upper layer, the above estimates are come close to the mean values obtained from year-long moored current observations of the SEMC at 23°S (17 Sv in GECCO and 20.6 Sv in Swallow et al., 1988), the Mozambique Current at 17°S (13.3 Sv in GECCO and 14 Sv in Ridderinkhof and de Ruijter, 2003) and the Agulhas Current at 32°S (78.4 Sv in GECCO and 75 Sv in Beal and Bryden, 1999).

The deep layer is confined within the depth range of 1300–3400 m, but varies geographically in its vertical extent and central depth (Fig. 5c). It covers the depth of NADW and its derivative, the
Lower Circumpolar Deep Water (LCDW), which are generally considered as the source of deep water in the Indian Ocean (Mantyla and Reid, 1995; Robbins and Toole, 1997; You, 1999). Constrained by the ocean bathymetry of the Indian Ocean, the northward inflow of NADW into the Indian Ocean at 32°S is confined to the western Indian Ocean. The deep transports recirculate in the Crozet Basin with a part of the transport flowing northward and the rest southward (Park et al., 1993; Mantyla and Reid, 1995). In contrast, the northward flow of LCDW into the Indian Ocean takes place only in the eastern half of the Indian Ocean along the Southeast Indian Ridge and Ninetyeast Ridge (Robbins and Toole, 1997; van Aken et al., 2004). Particular in the Perth Basin, the deep transports (and also bottom transports originating from AABW and LCDW) flow northward into the West Australian Basin and Central Indian Basin (Mantyla and Reid, 1995; McCarthy et al., 1997; Talley and Baringer, 1997; Warren and Johnson, 2002; Sloyan, 2006). In the model, the deep layer circulation is characterized by a small circulation and a relative small mass exchange with the adjacent oceans. Similar to the upper layer, the narrow western boundary contains the largest outflow of 3.3 Sv across 34°S in the deep layer, in which about 0.9 Sv re-circulates. Away from the western boundary, there is an obvious anti-cyclonic gyre near 60°E, in which 1.9 Sv enters the Indian Ocean between 60°E and 80°E and 1.0 Sv leaves west of 60°E. From 80°E towards the western coast of Australia, there is about 2.3 Sv inflow, most of which enters the Indian Ocean between 80°E and 100°E. A relative weak northward flow in the Mozambique Channel, which opposites to that of the upper and intermediate layer, feeds back to the south after bypassing
the northern tip of the Madagascar Basin, which is consistent with hydrographic observations (van Aken et al., 2004).

In the bottom layer the circulation is largely steered and controlled by bathymetry (Fig. 5d). Generally, there are two significant pathways of inflow. One is from the South Australian Basin, (about 2.1 Sv ± 4.0) to the West Australian Basin and then to the northern part of the Central Indian Basin via the Perth Basin and across the Ninetyeast Ridge with ∼1.6 Sv consistent with the work of Mantyla and Reid (1995). The other is in the Crozet Basin, to which about 1.2 Sv enters and circulates around the basin. It results in the outflow of 1 Sv and the rest flowing northward along the east coast of Madagascar Island, then enters the Somali Basin similar to the deep layer circulation. The strong western boundary current is absent in the bottom layer, and instead only 0.2 Sv exit via the Mozambique Basin.

The vertical mass transport in the interior Indian Ocean is a key to close the deep MOC of the Indian Ocean. Since the Indian Ocean has a strong seasonal monsoon reversal and lacks steady equatorial easterlies, no climatological equatorial upwelling exist; instead, upwelling mainly occurs along the coast of Somalia and Oman, the Arabian Peninsula and the east/west of the tip of India. The deep layer transfers the upwelling transports from the bottom layer to the above layer only with slight changes in its strength. However, the mass exchange between the upper layer and the intermediate layer is complex, since neither upwelling nor downwelling is dominant. The major downwelling occurs in the regions of Broken Plateau with 1 Sv, the Arabian Basin and Bay of Bengal (0.5 Sv and 0.3 Sv, respectively), the Mozambique Basin with 0.6 Sv, and the east of the Madagascar Basin with 0.3 Sv. In contrast, the upwelling is mainly along the African coast with 0.3 Sv in the north of the Somali Basin,
Fig. 5. Time-mean averaged circulation of the (a) upper, (b) intermediate, (c) deep and (d) bottom layer. Shown are only currents with magnitudes larger than 1 cm/s for the upper layer, 0.4 cm/s for the intermediate layer, 0.05 cm/s for the deep layer, and 0.1 cm/s for the bottom layer. The overlying grey arrows indicate the path along which currents follow and the shaded region indicates bottom topography with depths less than 3200 m.

0.3 Sv in the Madagascar Channel, 0.9 Sv in the western part of the Madagascar Basin in the band of 23°S–34°S, and with 0.5 Sv along the west coast of the Central Indian Basin.

Details about the inflow and outflow across the section 34°S in the GECCO solution are shown in Fig. 6. The upper layer is shallower than 1000 m and characterized by the broad northward inflow from 34°E to the west coast of Australia and the compensated narrow, strong southward western boundary current west of 34°E. The Agulhas Current has a width of the order of 600–700 km above 2000 m, reaching from the east coast of Africa to 34°E; its strength amounts to 60 Sv, 18.4 Sv, and 3.3 Sv southward for the upper, intermediate, and deep layer, respectively. From 34°E to the west coast of Australia, broad northward inflow carries 52.2 Sv and 14 Sv into the Indian Ocean in the upper and intermediate layer, while only 3.9 Sv and 2.1 Sv in the deep and bottom layer. Most of the inflow in the deep and bottom layers occurs in the eastern basin of the Indian Ocean (from 60°E to the west coast of Australia), consistent with the average circulation shown in Fig. 5. It results in a weaker lower limb of the deep MOC of 2.1 Sv via the South Australian Basin in the bottom layer.

4. Time-varying deep MOC of the Indian Ocean

According to Lee and Marotzke (1998), the MOC of the Indian Ocean displays large seasonal variations over the full depth of the water column with maximum variations near 10°S and 10°N, associated with the reversal of the monsoon over the northern Indian Ocean and changes of the easterlies over
the southern Indian Ocean. Although large seasonal variations of the MOC are confirmed by the GECCO synthesis, which shows large MOC variations basically south of 15°N, the meridional structure of the seasonal signal differs considerably (Fig. 7): during boreal winter, the northern Indian Ocean is dominated by two counter-rotating meridional overturning cells with upwelling near 10°S. The northern cell of 12 Sv covers the full depth of the model from 20°S to 15°N. Simultaneously, a meridional overturning cell of about 7 Sv resides in the region south of 20°S; it shows a core at around 1500 m depth and ranges from 500 m to the bottom at the southern boundary of the Indian Ocean. During boreal summer, the deep MOC is completely reversed.

We note that the transoceanic hydrographical section taken during November–December of 1987 was widely used in many box-inverse models. Due to the large seasonal variations in the strength of the deep MOC, it is therefore likely that those estimates are biased high, which partly explains why model results averaged over a full year are weaker in their time-mean deep MOC amplitude. More observations were used in inverse models to reach values of little more than 10 Sv of deep MOC (Ganachaud et al., 2000; McDonagh et al., 2008). However, since also these additional observations were taken during high MOC seasons, and observations taken during the period of low MOC are further to the north in areas were the overturning is in anti-phase to the circulation at 32°S, the estimates based on these observations are also likely to be biased high. According to the GECCO estimate, maximum values reach more than 11.5 Sv in some years during high MOC seasons. Considering the large spread responding to seasonal variations, MOC strengths in high MOC seasons in the deep and bottom layers are not far away from results of Ganachaud et al. (2000) and Robbins and Toole (1997) (Fig. 4). Moreover, with a value of 6.9 ± 1.5 Sv and 5.1 ± 1.5 Sv of maximum MOC during March and April in the bottom layer, respectively, our estimate is smaller than results of McDonagh et al. (2008) estimate, which is based on March and April measurements, but close within one standard deviation.
Fig. 7. The Indian Ocean meridional transport stream function (Sv) averaged over boreal (a) winter (December to February), and (b) summer (June to August).

Besides seasonal variations, the deep MOC also shows substantial interannual variability, which may provide further reasons for differences between estimates based on models or observations. As part of this, the southward upper layer transport, jointly with the upper layer ITF, is anomalously weak in the years of 1972/73, 1982/83, 1986/87, 1991/93 and 1997/98, coincident with IOD and ENSO events (Fig. 8). Two factors seem to contribute to this behaviour: firstly, the upper layer transport carries most of the net transport across 34°S and therefore shows a high correlation with the ITF for most of the study period (Fig. 9). Previous studies demonstrate a relation between ITF strength and ENSO, leading to larger ITF transports during a La Niña phase and smaller during an El Niño (Wang et al., 2004; England and Huang, 2005; Potemra and Schneider, 2007; Lan et al., 2009) but also indicate that this does not imply that ENSO dominates ITF variations, because of the role of the interannual variability in

Fig. 8. Normalized (by their maximum) 13-month running-mean time series of upper layer transport (bold solid), the upper layer ITF (dashed), IOD index (thin solid), and Nino3 index (shaded).
the Indian Ocean. Wunsch (2010) also found only minor ENSO contributions to ITF transports inferred from the ECCO-GODAE global estimate. Moreover, ITF may not determine the interannual variability of the upper layer transport of the Indian Ocean as larger variability of the upper layer transport than the ITF variability can be seen in the period 1970–1990 (Fig. 9). For the unfiltered times series, for which the seasonal cycle was removed, we find low correlation between ITF and the upper as well as bottom layer transports individually due to the impact of the MOC variability on the latter two transports, which shows little co-variability to the ITF. Since the overturning circulation is mainly affecting the upper and the bottom layer, its signal can be removed approximately from the upper layer transport by taking the sum of the upper and bottom transport. After doing so, a strong positive correlation (0.72) to the ITF is revealed, which indicates that the ITF mainly augments the upper layer transport.

The second process required to explain, why the upper layer transport variability is larger than the ITF variability, and why the bottom layer transport shows a strong negative correlation (−0.81) with the upper layer transports is apparently related to an overturning mode. The power spectra of the upper and bottom layer transports shows statistically significant peaks at interannual frequencies of 2-yrs and 3.5-yrs for the upper layer and periods of 1.4-yrs and 2-yrs for the bottom layer (Fig. 10a and b). The two-year periodicity present in both time series covaries with roughly a 12-month phase shift equivalent to an anti-correlation (Fig. 10c and d). The prominent two-year periodicity coincidences
with IOD variations and is associated with SST anomalies (Feng and Meyers, 2003), meridional heat transports variations in the equatorial region (Chirokova and Webster, 2006), and biennial oscillations described by Meehl and Arblaster (2002). It implies a connection between the deep MOC and tropical and subtropical ocean-atmosphere interactions, which are not well understood.

To further investigate a potential connection between tropical and subtropical ocean variations and the deep MOC, an extended EOF analysis was performed, using the annual mean MOC transports together with zonal (EEOFz) and meridional wind stress (EEOFm), respectively (Fig. 11). Spatially, the leading EOF mode of EEOFz is characterized by a basin-wide meridional overturning mode (Fig. 11a), which is nearly identical to the pattern of the 1st EOF mode from analyzing MOC transports alone (not shown). The mode highlights the connection between upper and bottom layer transports. The corresponding leading mode in EEOFm, although similar to the EOF mode described above, reveals strong vertical motion in the southern Indian Ocean (Fig. 11b). Moreover, both principle components (not shown) also have a close relation with that of the 1st mode of the MOC transports alone (0.82 for EEOFz and 0.62 for EEOFm, respectively). The zonal wind stress mode has a clear north–south dipole structure with opposite signs roughly on both sides of the 15°S section (Fig. 11c). Associated with meridional inflow or outflow resulting from Ekman transport, the structure is also favourable for upwelling or downwelling mostly in the southern Indian Ocean. The meridional wind stress mode
Our results suggest that Ekman transports and Ekman upwelling or downwelling play important roles in connecting the deep MOC to surface forcing. In addition, zonal wind stress along the equator appears to drive the upper ocean response and is closely related the Indian Ocean SST difference (IOD index). Alongshore wind stress off the Sumatra-Java coast also contribute to cold SST east of the equator via local upwelling, which is associated with the start of the life cycle of IOD events (Saji et al., 1999; Feng and Meyers, 2003). Moreover, the spatial pattern of wind stress (combining zonal and meridional wind stress mode) is similar to the September–October and November–December composites from Saji et al. (1999) based on six extreme events identified from the IOD index. All this reveals the close relation between wind stress modes and IOD, in which also wave dynamics induced by Ekman pumping may play a substantial role in the Indian dipole mode (Feng and Meyers, 2003; Le Blanc and Boulanger, 2001; Huang and Kinter, 2002). Therefore, Ekman dynamics and induced wave mechanism provide an intrinsic link between the deep MOC and the tropical and subtropical Indian Ocean on the interannual time-scale.

According to Schott et al. (2009), the lower limb of the deep MOC plays an important role in the bottom water mass exchange between the Indian Ocean and the Southern Ocean and it is more relevant for climate variability on longer time-scales. The upper layer transport across 34°S displays obvious

**Fig. 11.** Extended EOF analyses of meridional overturning with zonal (the 1st column) and meridional wind stress (the 2nd column). The 1st extended EOF mode for (top) meridional overturning and (bottom) zonal and meridional wind stress, account for 19%, and 21% of the total variance, respectively. The data is normalized by their standard deviations and the linear trend is removed before EOF analysis. Moreover, yearly averaged data is used here in order to highlight the interannual variation. (Fig. 11d) is associated with changes of meridional component of the easterlies, mainly along the diagonal of the southern Indian Ocean from north-west to south-east, and changes of the along-shore component near the Sumatra coast.
Table 1
Time mean and trends of volume transports (Sv) across 34°S in different layers and for different periods. Positive (negative) numbers mean northward (southward) flow.

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<tr>
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<tbody>
<tr>
<td>Upper</td>
<td>−7.6 ± 4.2</td>
<td>0.5</td>
<td>−8.4 ± 4.2</td>
<td>−2.0</td>
</tr>
<tr>
<td>Intermediate</td>
<td>−4.7 ± 1.5</td>
<td>−0.2</td>
<td>−4.0 ± 1.1</td>
<td>0.6</td>
</tr>
<tr>
<td>Deep</td>
<td>0.5 ± 2.1</td>
<td>0.</td>
<td>0.8 ± 1.8</td>
<td>0.8</td>
</tr>
<tr>
<td>Bottom</td>
<td>2.8 ± 4.1</td>
<td>0.2</td>
<td>1.5 ± 3.9</td>
<td>−1.4</td>
</tr>
<tr>
<td>ITF</td>
<td>−8.6 ± 3.5</td>
<td>0.6</td>
<td>−9.7 ± 3.6</td>
<td>−1.9</td>
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decadal to interdecadal variability and shows visible changes of the linear trends before and after 1980, which is also true for the bottom layer transport (Fig. 9). However, deep transports are subjected to slow adjustment processes, which inevitably affect long term trends also in ocean synthesis products. One has to keep in mind that the long term changes may therefore contain large errors. During 1960 through 1979, about 7.6 ± 4.2 Sv leave the Indian Ocean southward in the upper layer; in the period 1980–2001 the transport increases to 8.4 ± 4.2 Sv, while the bottom layer transport weakens from 2.8 ± 4.1 Sv to 1.5 ± 3.9 Sv (Table 1). Moreover on the long-term, trends of all layers transports during 1960–1979 remain more stable than the later period. For example, the strength of the upper layer transport decreases by 0.5 Sv in the period of 1960–1979, but increases by 2.0 Sv during the period of 1980–2001 simultaneously with the ITF increase (Table 1). During 1980 through 2001, substantial changes (−1.4 Sv) occur in the bottom layer, while only smaller changes of 0.6 Sv and 0.8 Sv occur in the intermediate and deep layer, respectively. Coincidently, the increase in the ITF matches that of the outflow via the upper layer and the changes of the intermediate and deep transports match those of the bottom layer.

Changes of the vertically integrated circulation between the two periods 1960–1979 and 1980–2001 are displayed for the four density classes in Fig. 12. The main changes within the upper layer in the northern Indian Ocean are in the equatorial circulation system, such as a stronger SEC, a stronger ITF and a correspondingly intensified SECC in the latter period. Those result in a stronger SEMC and an enhanced western boundary current, which is responsible for the increase of the upper layer transport during 1980 through 2001. For the intermediate layer, changes mainly occur at the two sides of section 34°S, which show a weaker subtropical gyre transport and a weaker western boundary current. Changes of the deep layer occur mainly along the west coast of the Somali and the Madagascar Basin, the Madagascar Channel, and the Crozet Basin. Major changes of the circulation in the bottom layer are confined to the slopes of the ocean bathymetry and lead to a decrease of the bottom inflow transport within the period of 1980–2001.

A complementary view on the transport changes is shown in Fig. 13 of the differences of the meridional velocity across 34°S. In the period 1980–2001, the intensified subtropical gyre results in a stronger compensated outflow of the western boundary current and an inflow of recirculation by strengthened Agulhas Current retroreflection. In addition, a weakened inflow occurring in the region of 70°E–80°E in the upper layer is noticeable. In the bottom layer, substantial differences are in the region of 60°E–77°E (the Crozet Basin).

5. Discussion and concluding remarks

Based on the GECCO synthesis over the period 1960 through 2001 an analysis of ocean volume transports in the Indian Ocean is performed with specific focus on the time-mean and time-varying characteristics of the deep MOC on different timescales. Results reveal that the time-mean integrated mass transport describes the strength of the lower limb of deep MOC with 2.1 Sv in the bottom layer which is roughly consistent with results from other ocean general circulation models and syntheses (Garternicht and Schott, 1997; Lee and Marotzke, 1998; Drijfhout and Garabato, 2008) within the range of 2–5 Sv, and a relatively strong outflow (12.3 Sv) of the upper and intermediate transport (above 1500 m) resulting from the transport of the ITF. The results are much weaker than estimates
Fig. 12. Differences of the integrated circulation of the (a) upper, (b) intermediate, (c) deep, and (d) bottom layer between 1960–1979 (P1) and 1980–2001 (P2). Only currents with speed larger than 0.1 cm/s are shown. The overlying grey arrows indicate the path along which currents follow and the shaded region indicates bottom topography with depths less than 3200 m.

Estimates of the inverse box model are based on inverse box models. Possible explanations of the large discrepancy are likely to result from the underlying assumptions and configurations of the inverse box model together with the sampling of ocean observations that concentrate on the seasons of high MOC.

Estimates of the inverse box model are based on the assumption of a steady state ocean. Different hydrographical sections of a specific time are, however, used as a constraint, in which the uncertainty of large seasonal variability is to a large extent introduced. As a result, estimates of inverse box models are highly dependent on the selected direct measurements. However, estimates of ocean general circulation models also suffer from a number of weakness, including open boundary conditions, horizontal and vertical resolutions, different surface heat and momentum fluxes products (SMCO1, Rintoul et al., 2010), and the prescribed excessively weak deep vertical mixing (Drijfhout and Garabato, 2008).

In our case, weak transports may firstly result from deficits of the GECCO model with respect to the bottom water formation, since GECCO seems to fall short in producing bottom water, resulting in a much weaker horizontal bottom transport and reduced upwelling in the tropical Atlantic, the Indian Ocean, and the entire Southern Ocean (Wang et al., 2010). Previous studies have already discussed a lack of bottom water in numerical simulations and suggested missing ice-ocean coupling and the lack of model resolution to be among the primary candidates for this deficit (Friocourt et al., 2005). Secondly, strong seasonal variations could explain why the time-mean deep MOC is much weaker than estimates based on observations taken during a season of strong transports.
In contrast to inverse box models, the GECCO synthesis provides time-varying, full-depth and three-dimensional global circulation fields. Our investigation reveals not only large seasonal and interannual variability, but also significant decadal to longer timescale variability in the Indian Ocean flow field:

- Large seasonal variations of deep MOC in the GECCO synthesis show the maximum strength of deep MOC at 34°S with $-7\text{ Sv}$ in boreal winter, and $3\text{ Sv}$ in summer. It results in a large spread of deep MOC among the seasons (Fig. 4). This may explain the discrepancy between estimates from the inverse box model and those from ocean general circulation models.

- The upper and bottom transports across 34°S show significant interannual variability with co-varying biennial variations, associated with the two-year timescale of IOD variations. By means of extended EOF analyses of MOC and the wind stress components, the relation of deep MOC to upper tropical and subtropical Indian Ocean is examined. The leading modes of zonal and meridional wind stress favour the basin-wide meridional overturning mode due to Ekman dynamics. Moreover, tropical zonal wind stress along the equator and alongshore wind stress off Sumatra-Java contribute to the evolution of IOD events. It reveals their close relations in which wave dynamics induced by Ekman pumping plays a substantial role in the Indian dipole mode (Le Blanc and Boulanger, 2001). Therefore, Ekman dynamics and the induced wave mechanism suggest an intrinsic link between the deep MOC and interannual variations of the tropical and subtropical Indian Ocean on the interannual time-scale.

- Variability on longer timescales is investigated for the period of 1960–1979 in comparison to 1980–2001, in which trends of all layer transports are found to differ before and after 1980. Trends remain relatively stable during 1960 through 1979, while larger in the period of 1980–2001, particularly for the upper and bottom layer. The comparison of the two periods shows that the intensified western boundary current resulting from an enhanced SEC is mainly responsible for the intensified upper layer transport. While in the intermediate layer, weakened western boundary current and
weakened inflow along the west coast of Australia lead to the weakened intermediate layer transport. Moreover, the changes in the deep and bottom layer transports are found to be mainly confined to the Somali Basin and the Madagascar Basin.

- A large part of the variability of the upper layer cannot be explained by the ITF variability, though most of exporting transports of the upper layer are from the ITF. However the correlation is recovered if the signal of the meridional overturning variability is removed, because the meridional overturning variability is largely uncorrelated with the ITF but strongly affects the variability of the upper and bottom layer.

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References


