Shallow Meridional Circulations in the Tropical Atmosphere

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ABSTRACT

A shallow meridional circulation (SMC) in the tropical atmosphere features a low-level (e.g., 700 hPa) flow that is in the opposite direction to the boundary layer monsoon or trade wind flow and is distinct from the meridional flow above. Representations of the SMC in three global reanalyses show both similarities and astonishing discrepancies. While the SMC over West Africa appears to be the strongest, it also exists over the eastern Atlantic and eastern Pacific Oceans, and over the Indian subcontinent, with different strength and structure. All SMCs undergo marked seasonal cycles. The SMCs are summarized into two types: one associated with the marine ITCZ and the other with the summer monsoon. The large-scale conditions for these two types of SMCs are similar: a strong meridional gradient in surface pressure linked to surface temperature distributions and an absence of deep moist convection. The processes responsible for these conditions are different for the two types of SMCs, as are their structures relative to moist convection, associated precipitation, and deep meridional overturning circulations. It is suggested that discrepancies among the representations of the SMC in the three global reanalyses stem from different treatment of physical parameterizations, especially for cumulus convection, in the models used for the data assimilation.

1. Introduction

A shallow meridional circulation (SMC) in the tropical atmosphere is a vertical–meridional overturning circulation similar to the commonly known deep circulation, except its return flow is in the lower troposphere (2–4 km). It is a local phenomenon. Even though its signature in the zonal mean is very weak, if identifiable at all, its amplitude can be comparable to local deep overturning circulations at certain longitudes. The key characteristic that distinguishes a SMC from a deep circulation is a maximum in the lower-troposphere meridional wind that is clearly distinct from the meridional wind aloft (Fig. 1a). Without this lower-tropospheric maximum in the meridional wind, the meridional overturning circulation is not considered as a SMC (Fig. 1b).

SMCs have been reported in several studies. An idealized simulation of the Hadley circulation (Schneider and Lindzen 1977) yielded, unexpectedly as a by-product, a zonal mean SMC in association with a prescribed meridional gradient in sea surface temperature (SST). In another idealized modeling study (Thorncroft and Blackburn 1999), a dominant SMC over West Africa was simulated in association with strong surface temperature gradients over the land as part of the heat low circulation. This SMC has been discerned in data assimilation products (e.g., Cadet and Nnoli 1987; Nicholson and Grist 2003). SMCs in other parts of the tropics have been shown in the divergent wind field of global reanalysis data (e.g., Tomas and Webster 1997). Trenberth et al. (2000) documented the global distribution and seasonal dependence of the SMC through an EOF analysis of the divergent wind field of a reanalysis dataset. SMCs have been observed in several in situ data of the meridional wind over the eastern Pacific Ocean (Zhang et al. 2004), West Africa (Fontaine and Janicot 1992; Parker et al., 2005; Zhang et al. 2006), and the western Pacific Ocean (Takayabu et al. 2006).

The existence of the SMC raises outstanding questions: What are the mechanisms for them to be shallow? What is their climatology? What are their roles in the tropical general circulation and climate? Several mecha-
nisms for the SMC have been proposed. Tomas and Webster (1997) explained the SMC in terms of cross-equatorial vorticity reversal. In a numerical simulation of Wang et al. (2005), radiative cooling at the top of marine stratus cloud south of the equator over the eastern Pacific Ocean is effective in enhancing the SMC, although not responsible for its existence.

Nolan et al. (2007) proposed that the primary mechanism for the SMC associated with the eastern Pacific ITCZ is a local response to regional gradients of surface temperature and surface pressure, when deep convection is absent. Typically in the ascent regions of the ITCZ surface temperatures are high and surface pressures are low, whereas the opposite is true in the nearby descent regions over the equatorial cold tongue and subtropics. Increased thickness owing to higher temperatures in the ITCZ region causes the pressure gradient between the ascent and descent regions to reverse above the boundary layer, and this, in turn, drives the shallow return flow out of the ITCZ. Nolan et al. showed that the essential features of the SMC could be reproduced from these considerations, both in a simple hydrostatic model and in an idealized, full-physics, tropical channel model of the ITCZ. The full-physics model also confirmed that the SMC is suppressed by deep convection and is strongest when deep convection is absent. The absence of deep convection in the ITCZ can be caused by synoptic-scale waves, intrusions of midtropospheric dry air (Zuidema et al. 2006), extratropical disturbances (Kiladis 1998), and an absence of a local source of instability (Sobel 2007).

To further understand the SMC and its climatic roles, it is important to document its climatology (seasonal cycle, interannual variability, and spatial distribution). Using in situ observations for this is next to impossible because of their sparsity in the tropics. For such exploration, one is tempted to use global reanalysis products for their continuous coverage in time and space and for their dynamic and thermodynamic consistency. Among the currently available global reanalyses, however, there are striking discrepancies with regard to their representations of the SMC in some regions (Fig. 12 in Zhang et al. 2006). Using a single reanalysis product to explore the climatology of the SMC can therefore result in biased or even misleading conclusions.

The present study is motivated by the need to explore the climatology of the SMC and the desire to learn from the discrepancies among global reanalyses. The purpose is to examine the climatology of the SMC using more than one global reanalysis product while documenting similarities, biases, and uncertainties in their representations of the SMC. Here, biases are discrepancies among the reanalysis products and in situ observations and uncertainties are discrepancies among the reanalysis products when in situ observations are unavailable. The reanalysis and observational data are described in section 2. Comparisons are first made among three reanalysis products and in situ observations in the SMC at limited locations in the tropics (section 3). Then the global distribution, regional structure, and seasonal cycle of the SMC are compared in detail for the three reanalysis datasets (section 4). In section 5, these comparisons are summarized and further discussion is presented, leading to conceptual models of two types of SMCs associated with the marine ITCZ and monsoon, respectively.

2. Data and methods

Three global reanalyses were used in this study. They are the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005), the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis 1 (hereafter NCEP1; Kalnay et al. 1996), and the NCEP–Department of Energy (DOE) Atmospheric Model Intercomparison Project (AMIP II) reanalysis (hereafter NCEP2;
Kanamitsu et al. 2002). All have the same horizontal resolution (2.5° × 2.5°) and pressure levels (1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 hPa). Twenty-two years (1980–2001) of reanalysis data were used for the comparisons. When the reanalyses were compared to in situ observations, their climatology was calculated over the period in which the in situ observations are available.

Long-term in situ observations are available for the purpose of this study only at limited locations. Two types of observations were used. Measurements using 915-MHz wind profilers at San Cristóbal, Galápagos (0.9°S, 89.7°W) and Christmas Island (2.8°N, 157.5°W) were available for 1995–2000. These data were used by Zhang et al. (2004) to detect the existence of the SMC over the tropical eastern Pacific Ocean. Their detailed descriptions are given by Gage et al. (1994). Atmospheric rawinsonde data from selected sites over West Africa have been used to document the SMC associated with the West African monsoon (Zhang et al. 2006). Data from three sites for 1980–2000 were used in this study: Abidjan (5.3°N, 4°W), Douala (4°N, 9.7°E), and Dakar (14.6°N, 17.5°W). The wind profiler data were collected from an open ocean environment. The sounding data were all taken from continental coastal sites. The reason for choosing in situ observations only from the eastern Pacific and West Africa, and not other parts of the tropics, is the availability of data as well as the robust signals of the SMC in these regions.

In addition, the version 2 rainfall data from the Global Precipitation Climatology Project (GPCP) were used to put the diagnoses of the SMC in the context of the ITCZ and monsoon rainbands. These data were produced by merging infrared and microwave satellite estimates of precipitation with rain gauge data from more than 6000 stations. Adler et al. (2003) provide detailed descriptions of this dataset. Monthly climatology on the 2.5° × 2.5° resolution for 1991–2000 was used in this study.

Two criteria were designed to identify the SMC: (i) the meridional wind at the 700-hPa level in opposite direction to that in the boundary layer and (ii) the meridional wind either decreasing in amplitude or changing direction between the 700- and 500-hPa levels. The first criterion ensures that the low-level (i.e., lower tropospheric) meridional wind is a “return flow” relative to the boundary layer meridional wind; the second distinguishes the lower-tropospheric meridional wind from those aloft (Fig. 1a). These criteria do not perfectly identify the SMC, but they serve well for the purpose of this study. They will be applied first subjectively based on visual inspection of figures and then objectively using digital wind data.

3. Comparisons between observations and reanalyses

Comparisons between meridional winds from the reanalyses and in situ observations at the two sites with wind profilers and the three sites with soundings are shown in Figs. 2–6. At Galápagos (0.9°S, 89.7°W), which is normally located south of the Pacific ITCZ, the wind profiler data (Fig. 2a) present a persistent boundary layer southerly flow. This boundary layer flow is mainly driven by the strong meridional SST gradient in the equatorial eastern Pacific (e.g., McGauley et al. 2004).

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1 Here, the boundary layer is loosely defined as the layer of high moisture content above the surface, whose top is characterized by a temperature inversion, and/or a strong vertical shear in wind speed or humidity, regardless of whether it is well mixed. Away from the ITCZ over the tropical eastern Pacific Ocean, a typical depth of this boundary layer is 1 km (e.g., McGauley et al. 2004).
2004), weakened only during boreal spring when the equatorial cold tongue is reduced in the region and a double ITCZ tends to develop (Zhang 2001). The southerly flow maximizes near the surface, quickly decreases with height, and reverses its direction into a weak northerly flow in the lower troposphere. This northerly flow, upon reaching its maximum in the lower troposphere, signifies a shallow return flow. This shallow return flow is strongest in boreal winter and weakest in late spring and early summer. Together with the southerly flow in the boundary layer, ascending motions in the ITCZ, and descending motions to the south, the pattern forms a weak SMC. All three reanalyses (Figs. 2b–d) reproduce the boundary layer southerly flow at this location, with correct seasonal cycle and depth. The boundary layer appears to be better mixed in the reanalyses, especially in ERA-40 and NCEP2, than in the observations. The boreal winter (December–January) maximum in the shallow northerly return flow is also present in all three reanalyses. However, NCEP1 fails to maintain the low-level northerly flow in the rest of the year.

At Christmas Island (2.8°N, 157.5°W), similar to the Galápagos, the wind profiler measurement exhibits a clear seasonal cycle in the boundary layer meridional wind (southerly), which is strongest and deepest during boreal summer and the opposite in boreal spring (Fig. 3). A northerly return flow is present at a higher level (600 hPa) than at the Galápagos, except during the spring. Two reanalyses (ERA-40 and NCEP2) reproduce the boundary layer southerlies and the northerly return flow in the summer, but all falsely produce a deep layer of strong boundary layer northerly flows in January–May. The ITCZ is always north of Christmas Island in reality, but in the reanalyses it is south of it in boreal winter, hence the reversal in the boundary layer meridional wind.

The West African region is also characterized by a prominent SMC (Parker et al. 2005; Zhang et al. 2006). At Abidjan (Fig. 4a), a city on the Guinea coast (5.3°N, 4°W), the sounding observations show a boundary layer onshore southerly flow persisting year-round. Its strength and depth undergo a marked seasonal cycle, both reaching peaks during the time of the summer monsoon. Above the onshore southerly flow in the boundary layer is an offshore northerly flow, which also undergoes a marked seasonal cycle. It is strongest during boreal spring and its level of maximum wind varies
between 800 hPa during boreal winter and 600 hPa during boreal summer. A similar seasonal cycle is seen from the sounding data at Douala, east of Abidjan, (4°N, 9.7°E: Fig. 5a), although its boundary layer southerly flow is much weaker. At these two locations, the three reanalyses reproduce the gross features of the seasonal cycle of both boundary layer and lower-tropospheric meridional flows. But the northerly return flows in the two NCEP reanalyses are much stronger than in ERA-40 and the sounding data, and the boundary layer southerly flows in ERA-40 and NCEP2 unrealistically collapse during boreal winter at Douala (Figs. 5b,d).

Dakar (14.6°N, 17.5°W) is the only sounding location chosen in this study that is constantly north of the ITCZ or monsoon rainband; therefore, a different perspective can be obtained from its observations (Fig. 6a). Here, the boundary layer is dominated by a strong northerly flow into the ITCZ, which is weakened only during boreal summer. Above this northerly flow is a shallow southerly return flow that becomes distinct from the meridional wind aloft only during boreal spring. Therefore, a SMC exists at this location only during boreal spring. This is so far the only observational evidence for a SMC north of the ITCZ. All three reanalyses reproduce this SMC (Figs. 6b–d).

These comparisons indicate that the SMCs in the reanalyses are not artifacts and that the reanalyses can even reproduce the observed gross features of the SMCs in most cases presented here, with an obvious exception at Christmas Island (Fig. 3). While the similarities between the reanalyses and in situ observations, and perhaps among the reanalyses themselves, may lend some confidence to the reliability of their representations of the SMC, the discrepancies are equally, if not more, interesting and intriguing. In the rest of this article, we explore to what extent the similarities among the reanalyses can be used to form hypotheses on the structure and dynamics of the SMC, and what we can learn from their discrepancies.

4. Global distribution

In this section, we compare the meridional circulations in the tropics represented by the three global reanalyses without reference to in situ observations, which are not available in most of the tropics. Through such comparisons we wish to systematically document...
similarities and discrepancies (which define uncertainties) among these reanalyses in their representations of the SMC.

a. Cross-equatorial flows

The seasonal and zonal distributions of the SMC across the equator are shown in Figs. 7–10 in which the mean meridional wind ($v$) along the equator is plotted for January, April, July, and October. A cross-equatorial SMC is indicated in these figures by the reversal of direction in the meridional wind from the boundary layer to the lower troposphere where the wind reaches its local maximum. To put discussions of the SMC in the perspective of the general circulation in the tropics, we will describe it alongside the better-known deep meridional overturning circulation, which is signified by opposite signs in the meridional wind between the boundary layer and the upper troposphere.

In boreal winter (e.g., January; Fig. 7), all three reanalyses show strong cross-equatorial northerlies ($v < 0$) in the boundary layer and southerlies ($v > 0$) in the upper troposphere in at least two-thirds of the tropics. They constitute the large-scale deep meridional overturning circulation. Their zonal mean defines the Hadley cell whose ascending branch is south of the equator. Exceptions are found over the eastern Pacific ($90^\circ$–$20^\circ$W) and eastern Atlantic Ocean/Gulf of Guinea ($20^\circ$E–$30^\circ$W) where the deep cross-equatorial meridional overturning circulations are in the opposite direction (southerlies in the boundary layer and northerlies aloft) from the rest of the tropics. This reflects the fact that, during boreal winter, the ITCZ in these regions and its associated large-scale ascending motion normally remain in the Northern Hemisphere (e.g., Waliser and Gautier 1993; Zhang 2001). The ascending
branch of the zonal mean Hadley cell in this season is located south of the equator largely due to monsoon convection (e.g., over Australia, South America, and Africa).

A SMC can be discerned clearly over the eastern Pacific Ocean in ERA-40 (Fig. 7a), where a distinct layer of northerlies (ν < 0) near the 700-hPa level exists above the boundary layer southerlies. The shallow northerly return flow is much weaker, if identifiable at all, in the other two reanalyses. Over the eastern Atlantic Ocean and Africa, on the other hand, all three reanalyses display strong signatures of an SMC. At certain longitudes, the surface and boundary layer cross-equatorial flows appear to have their return flows only in the lower troposphere.

In the transition season of boreal spring (e.g., April; Fig. 8), the upper-tropospheric meridional wind varies substantially with longitude and the zonal mean cross-equatorial flow is almost zero. In this season, the zonal mean Hadley circulation is nearly symmetric about the equator (Newell et al. 1972, p. 45; Peixoto and Oort 1991). The SMC over the eastern Pacific Ocean in ERA-40 becomes much weaker than in boreal winter. Over the eastern Atlantic Ocean and Africa, it remains strong in the two NCEP reanalyses but again substantially weakens in ERA-40.

In boreal summer (e.g., July; Fig. 9), when all large-scale deep convective centers are in the Northern Hemisphere, the deep meridional circulation emerges vigorously again throughout almost the entire tropics with a reversed sign from winter (except over the eastern Pacific and Atlantic Oceans) in all three reanalyses. The three regions of local maximum northerly flows in the upper troposphere and their counterparts of boundary layer southerly flows correspond well to the three major monsoon systems: the Asian monsoon, the West African monsoon, and the North American monsoon. The SMC over the eastern Pacific Ocean is present in ERA-40 and NCEP2, but not in NCEP1. This is the first, but not the only, example where one NCEP reanalysis agrees better with ERA-40 than with the other NCEP reanalysis. The SMC over the eastern Atlantic Ocean is seen only in ERA-40, over Africa only in NCEP1. Meanwhile, there is a weak signal of a SMC over the western Pacific Ocean (110°–160°E) in the two NCEP reanalyses. The coexistence of the SMCs and the deep meridional circulations is also apparent in an EOF analysis of the divergent wind field from one of the reanalyses (Trenberth et al. 2000).

The irregular pattern of the meridional circulation seen in boreal spring returns in fall (e.g., October; Fig. 10). The upper-tropospheric northerly flows are substantially weakened from the summer and, at certain longitudes, reverse their signs. The SMC over the eastern Pacific remains apparent in ERA-40 and NCEP2. Over the eastern Atlantic Ocean and Africa, the SMC is present in all three reanalyses but with different strengths.

The above comparisons clearly demonstrate that the SMC is not an isolated phenomenon in any specific reanalysis product. Meanwhile, the disagreement among the three reanalyses cautions on the use of any single one in the study of the SMC. The similarities and discrepancies among the three reanalyses in their representations of the SMCs are scrutinized next in regions where SMCs exist in at least one of the reanalyses.

b. Regional patterns

Based on Figs. 7–10, discrepancies among the three reanalyses notwithstanding, cross-equatorial SMCs exist clearly in regions of the eastern Pacific, eastern Atlantic, and West Africa. Next, we show examples of the vertical and horizontal structures of the SMCs in these regions. In all vertical–meridional cross-section figures in this section, the vertical wind is amplified 20 times to better illustrate the meridional overturning.

The first example is for the longitudes of West Africa (10°E–15°W) in April (Fig. 11). The contrast in the meridional circulations between ERA-40 and the two NCEP reanalyses is astonishing. An unmistakable SMC is present in both NCEP reanalyses (Figs. 11b and 11c), which extends between 10°N and 10°S with a strong cross-equatorial boundary layer southerly flow topped by a cross-equatorial northerly return flow at the 700-hPa level, as seen in Fig. 8. This SMC is very different from that derived from the divergent wind field in the region in July (Trenberth et al. 2000). In ERA-40, this SMC is completely missing over the ocean (south of 5°N, Fig. 11a). Instead, there are two deep overturning circulations on each side of the equator. In the mid- to upper troposphere, they are equatorially symmetric, sharing the same ascending branch. In the lower troposphere, however, the symmetry is broken by an ascending motion at 8°–10°N, roughly the same location of the ascending branches of the SMCs in the NCEP reanalyses. These lower-tropospheric ascents are collocated with the Saharan heat low, characterized by strong dry convection (Thorncroft and Blackburn 1999; Parker et al. 2005). At this time of the year, the major observed rainband is close to the Guinea coast but very weak (Fig. 13; also see Gu and Adler 2004). Figure 11 seems to suggest that all three reanalyses are able to reproduce the ascending motion and outflow associated with the Saharan heat low. But, deep convection in the rainband appears to be much stronger in ERA-40 than in the two NCEP reanalyses, as signaled by the ascending
motions near the equator. The suppression of convection in the NCEP reanalyses is consistent with the dry and warm advection by the northerly outflow of the Saharan heat low (Parker et al. 2005), but such advection is stronger in ERA-40. It is known that ERA-40 tends to produce excessive deep convective rainfall (A. Tompkins 2007, personal communication). Our results lead to a speculation that insensitivity of cumulus parameterization in the ERA-40 model to dry air might be a reason for the excessive rainfall over the Gulf of Guinea, hence the lack of a cross-equatorial SMC.

One important aspect of the SMCs is that they, as their counterparts of local deep meridional circulations, are not closed circulations because of the strong zonal flow. This can be seen from Fig. 12 in which horizontal wind vectors at the 700-hPa level are plotted together with observed precipitation. There is no qualitative difference among the three reanalyses in their representations of the horizontal circulation in this region. Over North and West Africa the large-scale circulation is anticyclonic, associated with the outflow of the Saharan heat low circulation (Thorncroft and Blackburn 1999). The associated northerlies penetrate through the latitude of the rainband along the Guinea coast and extend into the Southern Hemisphere in NCEP1 and NCEP2, where they are the low-level return flow of the cross-equatorial SMCs seen in Figs. 11b and 11c. In ERA-40, this northeasterly flow is confined to north of the equator.

To the west, over the Atlantic Ocean, the southerly component of the anticyclonic circulation is the low-level return flow of the SMC seen at Dakar (Fig. 6). The Saharan heat low may play a dominant role in generating the needed meridional gradient in surface pressure for the SMC over West Africa, the Gulf of Guinea, and perhaps the eastern Atlantic Ocean also. This suggests that the SMC might still exist in the tropical eastern Atlantic, even without the equatorial cold
tongue there. This is in contrast to the SMC over the tropical eastern Pacific where the strong meridional gradient in surface pressure, the main driver of the SMC there, is due to the equatorial cold tongue (McGauley et al. 2004).

The SMC over the central Atlantic Ocean (Fig. 13) is quite different from that over West Africa but also exhibits unexpected discrepancies among the three reanalyses, which may reveal something we can learn. Here both ERA-40 and NCEP2 produce strong ascending motions between 0° and 5°S (Figs. 13a and 13c). The strong ascending motions are the upward branches of two deep meridional cells, with the one across the equator stronger than the one in the Southern Hemisphere. The upward motion at these latitudes is much weaker in NCEP1 (Fig. 13b), making the deep meridional circulations less well defined, even though the boundary layer meridional convergence and upper-tropospheric divergence are obvious at the same latitudes as in the other two reanalyses. The latitudes of maximum boundary layer confluence of the meridional wind mark the location of the ITCZ in the reanalyses, which appears to be a little south of the ITCZ location indicated by the observed maximum precipitation (Fig. 12). In NCEP1 there is a clearly defined SMC north of the ITCZ. This SMC also exists in NCEP2 but is less discernible in ERA-40. The shallow southerly return flow of this SMC appears to be partially related to the anticyclonic circulation associated with the Saharan heat low (Fig. 12). It is unclear, however, whether the heat low circulation can extend into the open ocean without influences of the ITCZ, so this SMC might be a result of both the heat low over the land and the marine ITCZ. In addition, there is another SMC south of the ITCZ in NCEP2 with a much smaller meridional scale. There is no SMC south of the ITCZ in the other two reanalyses. Dual SMCs on both sides of the ITCZ can be seen in the divergent wind field of a global reanalysis (Trenberth et al. 2000).

The meridional–vertical structure of the SMC over the eastern Pacific in October in ERA-40 and NCEP2 is shown in Figs. 14a and 14c. The ITCZ is located at

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**Fig. 13.** As in Fig. 11 but averaged over the central Atlantic (15°–30°W): maximum vector amplitude 20.7 m s⁻¹.

**Fig. 14.** Vertical–meridional cross section of mean wind vectors averaged over the eastern Pacific (90°–160°W) during October (1991–2000) from (a) ERA-40, (b) NCEP1, and (c) NCEP2. The vertical velocity is amplified 20 times for better illustration. Maximum vector amplitude is 13.1 m s⁻¹.
5°–10°N, as indicated by the boundary layer meridional confluence and strong ascending motions. The deep overturning circulations are well defined on both sides of the ITCZ. The SMC south of the ITCZ extends across the equator to about 10°S. The shallow northerly return flows of the SMC are part of the northeasterly wind covering a broad region south of the ITCZ where the southeasterly trade wind dominates in the boundary layer (not shown). Different from the SMCs seen before, the shallow northerly return flows occupy a deeper layer (700–500 hPa) of descending motion. There is no obvious SCM in NCEP1 over the eastern Pacific in this month (Fig. 14b).

Figures 11–14 were presented to highlight the most significant discrepancies among the reanalyses. But their key similarities can also be seen, as illustrated in the next section.

c. Seasonal cycle

The seasonal cycle of the SMC has been partially shown in Figs. 2–6 at single points without comparison to the deep meridional circulation and in Figs. 7–10 along the equator. Here it is discussed in comparison to the deep meridional circulation in selected regions where either cross-equatorial or off-equatorial SMC is particularly strong in at least one of the reanalyses. The seasonal cycle of meridional wind at the equator over the longitudes of West Africa (10°E–15°W) is shown in Fig. 15. The local deep meridional overturning circulation associated with the West African monsoon, with a strong boundary layer southerly cross-equatorial flow toward land and a northerly cross-equatorial return flow toward the ocean in the upper troposphere, is the dominant feature of the seasonal cycle in all three reanalyses. Meanwhile, the SMC, signified by the weaker but evident northerly return flow in the lower troposphere, exists also in all three reanalyses but with different seasonal variability. Its largest amplitude occurs during boreal spring in both NCEP reanalyses, before the peak of the local deep meridional circulation in boreal summer (Figs. 15b and 15c). In NCEP1 it almost disappears during boreal summer, whereas it persists through the entire year in NCEP2. The SMC exhibits a unique semiannual cycle in ERA-40 (Fig. 15a). In all three reanalyses, the height of the SMC return flow varies seasonally with the depth of the southerly monsoon flow, which is greatest during the monsoon peak season, as observed from in situ sounding data (Zhang et al. 2006).

Over the equatorial eastern Pacific (90°–160°W), the SMC in ERA-40 is established in July–January, after the peak of the local deep meridional circulation in May–July (Fig. 16a). This is in contrast with the sea-
sonal cycle over the West African longitudes. The SMC is much weaker, if identifiable at all, in the two NCEP reanalyses (Figs. 16b and 16c).

In addition to the cross-equatorial SMCs on the southern side of the ITCZ or monsoon rainband, SMCs also exist on the northern side, as suggested by Fig. 13. In all three reanalyses, the SMC north of the Atlantic ITCZ (15°N, 15°–30°W; Fig. 17) is indicated by a low-level southerly return flow (away from the ITCZ) paired to the boundary layer northerly inflow (toward the ITCZ). It is strongest in boreal spring and disappears in summer. The maximum of the return flow tends to be at a higher level (500 hPa) in ERA-40 than in the two NCEP reanalyses (700 hPa).

There is an SMC north of the Indian monsoon rainband (15°N) over 70°–90°E, especially during the monsoon onset (May) in all three reanalyses (Fig. 18). It is similar to the one over West Africa (Fig. 11) but is located much farther north. This Indian SMC is strongest and most persistent in NCEP2 and weakest in ERA-40. It appears to be related to the anticyclonic circulation over northern India (Fig. 12). It is possible that this SMC is related to a heat low over northern India, similar to the one over West Africa. The difference is that the Indian monsoon rainfall advances northward to the latitude of this heat low and obliterates it along with the SMC during the peak season of the monsoon in ERA-40 and NCEP1. This northward migration of the monsoon rainband is signified by the reversal of the sign in the deep meridional overturning circulation during June–August. This relationship among the SMC, heat low, and monsoon precipitation underlines the essential role of the heat low in the SMC associated with the monsoon.

The previous discussions are mainly based on subjective judgment as to where an SMC exists. To provide an objective description of the SMC in the three reanalyses, including its seasonal cycle and zonal distribution, the two criteria introduced in section 2 were quantitatively checked against the wind data. Equatorial meridional winds at the 700-hPa level, plotted in Fig. 19 as a function of month and longitude along the equator, are shaded when these two criteria are satisfied.

With different strengths, all three reanalyses present well-defined cross-equatorial SMCs over West Africa (0°–20°E) whose peaks are all in boreal winter and spring. Other SMCs, for example, over the central/ western Atlantic Ocean, South America, and the eastern and western Pacific Ocean, all differ substantially in both strength and seasonality among the three reanalyses. It is noted that all cross-equatorial SMCs are in the same direction as the northerly return flow. The strong southerlies off the Somali coast that reached peaks dur-
ing July and August do not form an SMC but are part of the deep flow of the summer Indian monsoon.

When the two SMC criteria are applied at 15°N (Fig. 20), they reveal SMCs north of the ITCZ and the monsoon rainband. These SMCs have their shallow southerly return flow ($v > 0$) poleward of the ITCZ, normally located at 5°–10°N, and their boundary layer northerlies into the ITCZ (Figs. 13, 17, and 18). Such SMCs over the Atlantic Ocean (around 30°W), in the direction opposite to the one south of the ITCZ, is consistently present in all three reanalyses during most of the year except boreal summer (an example is presented in Fig. 13). A much weaker and less consistent SMC can be found in the eastern Pacific in the two NCEP reanalyses. The strong negative meridional wind with shading in Fig. 20 over the Indian Ocean (80°–90°E) during boreal spring in all reanalyses indicate the SMC seen in Fig. 18. These SMCs show no fundamental difference from those identified with the ITCZ located at the equator (Nolan et al. 2007).

An SMC also exists over South America in boreal summer in the divergent wind field of a global reanalysis (Trenberth et al. 2000). It is evident in the total wind field only from NCEP1 (Fig. 19b).

5. Summary and discussion

Representations of the SMC in three global reanalyses (ERA-40, NCEP1, and NCEP2) have been compared to limited in situ sounding observations. The SMC is too strong in ERA-40 but too weak in NCEP1 over the eastern Pacific, whereas it is too strong in NCEP1 and NCEP2 but too weak in ERA-40 over West Africa. While these biases can be large at certain locations, the seasonal cycle in the SMC represented by the reanalyses is consistent with observations in most cases. These comparisons provide some confidence that SMCs in the reanalyses are not artifacts and a global survey of the SMC in the tropics based solely on the reanalyses may provide useful knowledge on this phenomenon not available from in situ observations.

The global distribution, structure, and seasonal cycle of the SMC in the three reanalyses have then been documented. Uncertainties in the SMC, defined as discrepancies among its representations in the reanalyses, can nevertheless be substantial over open oceans where no observations are available to constrain the reanalysis.
products. There is no consistent pattern in the uncer-
tainties. The two NCEP reanalyses do not always agree
with each other more than with ERA-40, and the two
more recent reanalyses (ERA-40 and NCEP2) do not
always agree with each other more than with the ear-
est one (NCEP1). All three reanalyses differ from
each other over the Atlantic Ocean during April (Fig.
13). Large uncertainty is also found over West Africa in
April when the shallow northerly return flow of the
SMC is very strong in the two NCEP reanalysis but
almost fails to appear in ERA-40 (Fig. 11). Over the
eastern Pacific Ocean, the SMC in October exists in
both ERA-40 and NCEP2 but not in NCEP1 (Fig. 14).

These and other discrepancies caution against using a
single reanalysis product to investigate the SMC and
perhaps other tropical phenomena. Consistency among
different reanalyses, even though not a guarantee of
correctness, reduces the uncertainty to a level at which
our view of the SMC is not obscured by known contra-
dictions. Such consistency can indeed be found among
the three reanalyses:

(i) the strongest SMC is over West Africa in boreal
spring,
(ii) all SMCs undergo marked seasonal cycles,
(iii) SMCs can exist on either side of the marine ITCZ,
and
(iv) an SMC can exist in association with either the
marine ITCZ or monsoon.

The SMC makes an interesting comparison to the
depth meridional overturning circulation. The deep me-
ridional circulation, more consistently represented by
the three reanalyses than the SMC, is most robust in
boreal summer (Fig. 9) but varies considerably with
longitude in the other seasons (Figs. 7, 8, and 10). The
SMC is always a local phenomenon with its peak in
different seasons at different longitudes (spring over
West Africa, summer over the Atlantic Ocean, and fall
over the eastern Pacific). At longitudes where deep
convection associated with the summer monsoons mi-
grates between the hemispheres, the deep meridional
circulation reverses its direction between the solstitial
seasons. There is no such seasonal reversal in the di-
rection of the SMC (and the deep meridional circula-
tion associated with the marine ITCZ over the open
oceans, e.g., the central and eastern Pacific and Atlan-
tic). The seasonal cycle in the SMC is seen mainly in its
strength and vertical structure. The differences be-
tween the deep and shallow meridional circulations
suggest that they are associated with fundamentally dif-
ferent processes. They both need a strong meridional
gradient in surface pressure to drive the lower-branch
flow in the boundary layer. They are distinguished from
each other by the depth of their associated convection.

The EOF analysis by Trenberth et al. (2000) on di-
vergent wind clearly shows the divergent component of
the SMC. It is interesting to ponder whether and how
much rotational wind may also contribute to the SMC.
A substantial contribution to the SMC from rotational
wind would have a different dynamic interpretation
from the association with the depth of convective heat-
ing. To address this question, we calculated both rota-
tional and divergent components of the wind fields. Fig-
ures 21 and 22 present an example for April. It is evi-
dent that rotational wind, while strong, does not
contribute much to the SMC (i.e., northerly flows at the
700-hPa level) over West Africa and India. Similar re-
results are also found for the SMC in the rest of the
tropics and the rest of the year (not shown). These two
figures suggest that the SMC is perhaps not limited to
the meridional direction but can be three-dimensional;
not, for example, the divergent flow away from the
African continent on the western and eastern sides of
the continent. This pattern also confirms our thinking
that the SMC is mainly determined by the vertical
structure of convective heating, which is elaborated be-
low. They also suggest that any discrepancies between
the SMC, documented here and in Trenberth et al.
(2000), would be more likely to come from EOF modes
versus total wind fields than from contributions by di-
vergent versus nondivergent components.

The comparison between SMCs associated with the
marine ITCZ and the monsoon is equally interesting.
Both exist under similar large-scale conditions: a strong
meridional gradient in surface pressure as the main
driver of the boundary layer circulation and an absence
of deep convection usually associated with deep over-
turning circulations. The processes that provide these
conditions are different for these two types of SMCs,
such as their structures relative to precipitation and the
depth overturning circulations. As discussed previously,
over the tropical eastern Pacific the meridional gradient
in surface pressure is maintained mainly by the equa-
torial cold tongue (Lindzen and Nigam 1987; McGauley
et al. 2004). Air–sea interaction is instrumental to the
existence of the equatorial cold tongue (e.g., Pike 1971;
Xie and Philander 1994). There, the ascending branches
of the SMC and the deep overturning circulation coin-
cide in the ITCZ. In the mean, they appear to coexist.
At any instant in time, the depth of the overturning
circulation is determined by the vertical structure of the
diabatic heating profile in the ITCZ (Fig. 23). The vari-
ability in the depth, strength, and frequency of convec-
tion in the ITCZ is constantly modulated by several
factors (see section 1). The structure of this “ITCZ SMC” is depicted in the schematic diagram of Fig. 24a. It is a different story for the “monsoon SMC”: Over West Africa, for example, the meridional gradient in surface pressure needed for the SMC is maintained mainly by heating of the land surface. Air–land interaction is essential to this, and an equatorial cold tongue to the south may not be necessary. Because the maxima in surface potential temperature and equivalent potential temperature do not coincide over the land (Thornicroft and Blackburn 1999), neither do the ascending branches of the SMC and deep overturning circulation associated with monsoon rainfall. As a consequence, the ascent branch of the SMC is located north of the monsoon rainband and its shallow return flow penetrates through the latitude of monsoon rainfall. These features of the monsoon SMC are distinct from the ITCZ SMC. As long as the monsoon rainband does not advance northward far enough to reach the heat low, dry convection would keep its overturning circulation shallow. This monsoon SMC over West Africa is sketched in Fig. 24b. It applies also to the SMC over

Fig. 21. Streamfunction contours \((10^6 \text{ m}^2\text{s}^{-1})\), vorticity (colors, \(10^{-6}\) s\(^{-1}\)), and rotational wind vectors at the 700-hPa level during April (1991–2000) from (a) ERA-40, (b) NCEP1, and (c) NCEP2.
India, but only for the period before monsoon rainfall advances northward and eliminates the heat low. It is noted that the dislocation of the ascents associated with the monsoon SMC and the deep meridional circulation is not seen from the EOF analysis of Trenberth et al. (2000).

The two types of SMCs may have different implications for tropical dynamics and climate. The low-level outflow of the ITCZ SMC transports moisture away from the ITCZ, serving as a sink or “leak” to its water vapor budget. When deep convection in the ITCZ is weak and the SMC is strong, this leak can be as large as 50% of the moisture into the ITCZ by the boundary layer convergent flow (Nolan et al. 2007). The moisture transported by the shallow return flow of the SMC may affect trade cumulus and stratus outside the ITCZ. The large vertical wind shear across the top of the boundary layer when the SMC is strong may provide extra turbulent mixing needed for the momentum balance in the boundary layer (Stevens et al. 2002; McGauley et al. 2004).

The distinct two-mode nature of the meridional over-
turning circulation associated with the ITCZ, with a deep and a shallow circulation, is especially intriguing. This is obvious in the EOF analysis of the divergent wind field by Trenberth et al. (2000) as well as in the total wind field shown in this study. This implies two distinct modes in diabatic heating profiles. Such two-mode heating profiles have been produced by an idealized model for the SMC (Nolan et al. 2007), and they may indeed exist in reality (Shige et al. 2007). This seems to be contradictory to the idealized simulation of Schneider and Lindzen (1977) in which only a deep diabatic heating profile was prescribed but both a deep and a shallow meridional overturning circulation were generated. However, in their model, the circulation does not feed back to diabatic heating. The boundary layer meridional wind forced by the prescribed gradient in SST was essentially a dry flow and resulted in a SMC exactly because of the sea-breeze mechanism of Nolan et al. (2007). It remains an open question, however, whether the two distinct vertical modes in the meridional overturning circulation can be explained solely in terms of large-scale fluid dynamics in the tropics.

The monsoon SMC may play an active role in the seasonal evolution of monsoon rainfall. Over West Africa, for example, the shallow return flow of the SMC is likely to advect dry, warm, and perhaps dusty, air from the heat low into the monsoon rainband (Parker et al. 2005). It is known that dry air in the lower to mid troposphere is detrimental to deep convection (e.g.,

**Fig. 23.** Vertical profiles of mean and composite moist heating and meridional velocities from idealized simulations of the ITCZ, Hadley circulation, and the shallow circulation by a regional mesoscale model: (a) diabatic heating from both resolved and parameterized convection 30 km north of the equator, for the time-zonal mean circulation (solid line), the strong shallow return flow composite (dashed line), and the weak shallow return flow composite (dashed– dotted line); (b) meridional velocities 400 km north of the equator for the mean circulation (solid line), strong shallow return flow composite (dashed line), and weak shallow return flow composite (dashed–dotted line). The strong and weak shallow return flow composites were determined from a limited-region average over periods when the shallow return flow was one standard deviation above and below mean strength, respectively (from Nolan et al. 2007).

**Fig. 24.** Schematic diagrams of the shallow meridional circulations (solid arrows) associated with (a) the ITCZ and (b) the monsoon rainband (represented by cloud symbols). The approximate locations of the equator and heat low are marked. The boundary layer is shaded. The deep meridional circulations are indicated by open arrows. The arrows depict only the direction of the circulations, not their strength. Surface convergence is expected to occur below vertical motions.
The dry, warm, and dusty advection by the shallow return flow of the monsoon SMC to the latitude of the monsoon rainband may inhibit its northward migration during the early stages of the monsoon season. When the instability inland is built up to the degree that deep convection can no longer be suppressed, the monsoon rainband suddenly occurs there in June, giving an impression of a northward jump (e.g., Sultan and Janicot 2000; Gu and Adler 2004). If this is the case, correctly reproducing the SMC by forecast models would be critical to predicting the seasonal monsoon evolution. In boreal winter, on the other hand, satellite observations show that aerosols released from biomass burning along the Guinea coast spread southward over the Gulf of Guinea. The perpetual boundary layer wind there is southerly. The southward advection of the aerosols must be advected by the shallow northerly return flow of the SMC.

The similarities and discrepancies among the three reanalyses, as well as between them and in situ observations, in their presentations of the SMC are very telling. They suggest that the fundamental dynamic ingredients for the SMC can be produced by the reanalysis models, which produce SMCs over the open ocean without much direct input from observations. The detailed mechanisms, however, may be sensitively related to physical processes that have to be represented by parameterizations, which can differ substantially among the models. One such process likely includes the depth, intensity, and frequency of moist convection. Based on the SMC mechanism proposed by Nolan et al. (2007), namely that the existence of an SMC depends on weakened or absent deep convection in the ITCZ, a model that overproduces deep convection would tend to underestimate the strength of the SMC there. Many current state-of-the-art coupled global climate models are completely incapable of reproducing the SMC over West Africa (Fig. 6 in Cook and Vizy 2006). This suggests that they might produce excessive deep convection. Although a cumulus scheme determines the vertical structure of convection, and hence the depth of the overturning circulation, a land surface parameterization determines the surface heating efficiency and the strength of the heat low, and a boundary layer parameterization determines the strength of inflow in a given surface pressure field. The existence and strength of the SMC in a model depends, therefore, on all of these, and perhaps other, parameterizations.

In conclusion, their large discrepancies notwithstanding, the three global reanalyses reveal that the SMC is a robust local phenomenon associated with either the marine ITCZ or the summer monsoon. Taking the SMC into consideration may help improve our understanding of the dynamics of the ITCZ, monsoon, and tropical circulation in general. The identified biases and uncertainties in the presentations of the SMC call for more in situ observations to further validate and improve the reanalyses.

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