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Deep-Sea Research II 52 (2005) 409–428

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DEEP-SEA RESEARCH  
PART II

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# Impact of the equatorial deep jets on estimates of zonal transports in the Atlantic

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Received 9 December 2003

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## Abstract

The structure and variability of the zonal equatorial flow in the Atlantic is studied on the basis of velocity profiles obtained with lowered Acoustic Doppler Current Profilers during multiple surveys. The vertical extent of the zonal currents is found to vary considerably. It can be as small as 100 m or as large as 1000 m. In the Atlantic, vertical scales of 400–600 m have been associated with the equatorial deep jets (they are also frequently called deep jets or stacked jets). Typical amplitudes of the zonal velocity are about  $20 \text{ cm s}^{-1}$ . An analysis of quasi-synoptic surveys indicates that the zonal extent of most jets is likely to be at least  $27^\circ$ . They can rise or deepen from west to east, although the deepening was observed more often and is often more pronounced. The west to east deepening can be as large as  $320 \text{ m}/10^\circ$ . Basin-wide mean depth changes of the jets are mostly on the order of  $50 \text{ m}/10^\circ$ , and the largest depth changes are typically observed between  $35^\circ$  and  $23^\circ \text{W}$ . The existence of these changes indicates that vertically propagating, equatorially trapped, waves might be one cause for the jet structure. However, the dependence of the slope on the longitude indicates that other processes must be involved as well.

The typical vertical extent of the jets is small enough to result in several direction changes of the zonal flow in the Antarctic Intermediate Water (AAIW) and the North Atlantic Deep Water (NADW) layer. From transport estimates for 14 meridional sections it is found that the transport for the westward component of the flow within the AAIW layer (500–1000 m) can be as large as  $-24 \text{ Sv}$  ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) within  $1^\circ$  of the equator. For the eastward component of the flow in the AAIW layer the transport can be as large as  $8 \text{ Sv}$ . Adding the transport components for each section results in a range of total AAIW transports from  $-24$  to  $7 \text{ Sv}$ . This suggests that the annual mean transport of AAIW is westward. The only months with eastward total transports are June and July. This is consistent with earlier Lagrangian and some other observations that indicated that the AAIW flow along the equator is governed by an annual cycle. In the NADW layer (1200–3900 m) the transport for the westward (eastward) flow can be as large as  $-25 \text{ Sv}$  ( $23 \text{ Sv}$ ) within  $1^\circ$  of the equator. This results in a range of total NADW transports from  $-10$  to  $18 \text{ Sv}$ . The variations of the total

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transports of AAIW and NADW are anti-correlated, with a correlation coefficient of  $-0.86$ . Since only eight sections reached deep enough to allow transport estimates in the NADW layer it is more difficult to come to a conclusion about the mean transport in this layer than for the transport in the AAIW layer (for the latter layer 14 sections were available). Nevertheless, the obtained estimates suggest that the total NADW transport may be eastward.

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## 1. Introduction

The zonal flow within about  $1^\circ$  of the equator is characterized by equatorial deep jets of opposing zonal velocity in the Atlantic (e.g., Ponte et al., 1990), Indian (e.g., Luyten and Swallow, 1976) and Pacific Ocean (e.g., Eriksen, 1981). The equatorial deep jets are also frequently called deep jets, stacked jets, or simply jets. We will use the word jet throughout this study. Gouriou et al. (1999, 2001) reported that the vertical scale of the jets in the Atlantic is 400–600 m, which is about 200 m larger than in the Pacific. Similar vertical scales were derived by Johnson and Zhang (2003) on the basis of hydrographic observations. For jets the typical peak-to-peak amplitude of the zonal velocity is on the order of  $20 \text{ cm s}^{-1}$ . Less is known about the horizontal scales of these jets. Gouriou et al. (1999, 2001) found indications that the zonal scale may be similar to the  $10\text{--}15^\circ$  observed in the Pacific and in the Indian Ocean (based on non-synoptic data), and may even be as large as  $25^\circ$ . Gouriou et al. (2001) suggested that the jets are not established simultaneously over the whole water column, since they observed that the zonal coherence of the jets depended on the depth in the summer of 1999.

It is known that the structure of the jets varies significantly over time, but the time scales of this variability are only poorly understood, since contradictory observations exist. Some support seasonal reversals of the zonal flow below the Equatorial Undercurrent, for example: time series from current meter moorings in the western (Weisberg et al., 1980; Weisberg, 1980) and eastern Atlantic (Thierry, 2000), and the Pacific (Eriksen, 1981; 1985); Lagrangian observations at intermediate depth in the Atlantic (Richardson and Schmitz, 1993; Boebel et al., 1999; Molinari et al., 1999; Schmid et al., 2001); and evidence from

velocity profiles in the Atlantic (Schott et al., 1998; Gouriou et al., 1999; Schott et al., 2003). Due to the nature of these observations it is hard to say if the seasonal reversals are due to the jets or due to currents with larger vertical scales than the jets. In addition, other observations do not support seasonal reversals, for example: observations in the Pacific do not show seasonal reversals of the jets or the flow, but the flow between 500 and 1700 m shifts from predominantly westward in April–July/1982 to predominantly eastward in April–July/1983 indicating that a large interannual variability may exist (Pegasus profiles from 21 cruises; Muench et al., 1994). Observations in the western Atlantic in October 1992 to May 1994 also do not show signs for seasonal reversals (four current meters between 1300 and 1900 m and Lowered Acoustic Doppler Current Profiler sections; Send et al., 2002).

In a modelling study Böning and Schott (1993) showed seasonal reversals of the flow below about 400 m, which they linked to seasonal changes of the equatorial wind. They noted that the vertical scale of the flow in their model is larger than the vertical scale of the observed jets, which may be primarily due to a vertical grid resolution that is too large to resolve the jets (250 m below a depth of 500 m). Therefore, the model cannot be used to determine if the jets reverse on seasonal to interannual time scales.

Many studies have focused on finding dynamical explanations for the jets. Superpositions of Rossby and Kelvin waves with long periods can give rise to jet-like flow patterns (e.g., McCreary, 1984). One problem with this hypothesis is that the meridional extent of the resulting jets is larger than observed ( $1.5^\circ$  instead of  $1^\circ$ ). Muench et al. (1994) showed that the addition of higher-frequency variability (e.g., Rossby-gravity waves with periods of 30–90 days) makes it possible to derive a

meridional extent that matches the observations. Other problems with the linear wave theory are that: (1) it predicts that the jets are symmetrical around the equator (e.g., McCreary, 1984; Pedlosky, 2003), whereas the maximum velocity cores of the jets have been observed to be at an off-equatorial position in some sections (e.g., Gouriou et al., 2001; Johnson et al., 2002). (2) This theory cannot explain how the jets are maintained for extended periods of time. As a mechanism for this, Muench and Kunze (2000) suggested that an ambient internal wave field which deposits momentum fluxes at critical layers within the deep jets can counteract the damping of the linear waves.

The generation of the jets also remains controversial. Suggested remote forcing mechanisms are: (1) surface forcing by the wind field (e.g., Wunsch, 1977). This is unlikely since the vertical group velocity is too small to generate and maintain the jets by surface forcing, even for weak dissipation (Muench and Kunze, 2000). Another problem is that a strong EUC can act as a barrier to vertically propagating waves. (2) Time-dependent boundary forcing, due to variability of the boundary currents (e.g., Ponte, 1989). (3) A specified zonal inflow from the boundary at the depth of the jets (Wang et al., 1994). Possibilities (2) and (3) are unlikely because the vertical scales of the jets do not match the vertical structure of the boundary currents and since there is no mechanism that can explain such a boundary forcing (Pedlosky, 2003). As an alternative to surface forcing Pedlosky (2003) suggested buoyancy forcing at the base of the mixed layer. A different line of thought suggests that local forcing by equatorial inertial instability (horizontal shear of the flow) can give rise to zonal flows of finite amplitude that have jet-like characteristics (Hua et al., 1997). If the flow is westward they have the maximum velocity off the equator, at the dynamical equator, which agrees better with the observations (e.g., Gouriou et al., 2001; Johnson et al., 2002).

In this study velocity sections obtained in the equatorial Atlantic over a period of 13 years will be analyzed. While many of the sections have been discussed before, a joint analysis of all these sections was never performed. It has to be

cautioned, that the available extensive data set is not sufficient for a full dynamical analysis of the observed flow structures. However, some interesting insights can be gained through its analysis. One purpose of this study is to describe the zonal and vertical structures of the flow (Section 3.1). The temporal variability will be discussed on the basis of these observations and earlier results (Section 3.2). The implications of this variability for estimates of the transports of Antarctic Intermediate Water (AAIW) and North Atlantic Deep Water (NADW) along the equator will be addressed (Section 3.3). In Section 4 a comparison of the characteristics of the jets found in the observations with the flow structures expected from the various theories about the jets will be presented in an effort to increase the understanding of the jets. In this section the transport estimates will be discussed in conjunction with earlier results.

## 2. Data and methods

The data set used in this study consists of Eulerian velocities from Lowered Acoustic Doppler Current Profiler (LADCP) and Pegasus (only Cither 1) measurements in the tropical Atlantic obtained during 14 cruises from 1990 to 2002 (Fig. 1, Table 1). During most of the cruises cross-equatorial sections were obtained. The only exception is the cruise performed in the summer of 1997 on board of the R.V. *Seward Johnson* during which a zonal section along the equator from 11° to 38°W was completed with a station spacing between 2° and 7° (Fig. 2A). During the course of five cruises cross-equatorial sections were obtained at two or three different longitudes. Ten of the cruises crossed the equator at about 35°W (Table 1). Another profile on the equator at 35°W was obtained in July 1997 (R.V. *Seward Johnson*). Other longitudes with repeated equatorial profiles include 23°W (3 profiles) and 10–12.5°W (4 profiles).

Depth-averaged differences between LADCP velocities and Pegasus velocities are about  $1 \text{ cm s}^{-1}$ , but below 1000 m, the uncertainties of LADCP velocities are  $3\text{--}5 \text{ cm s}^{-1}$  (Hacker et al.,

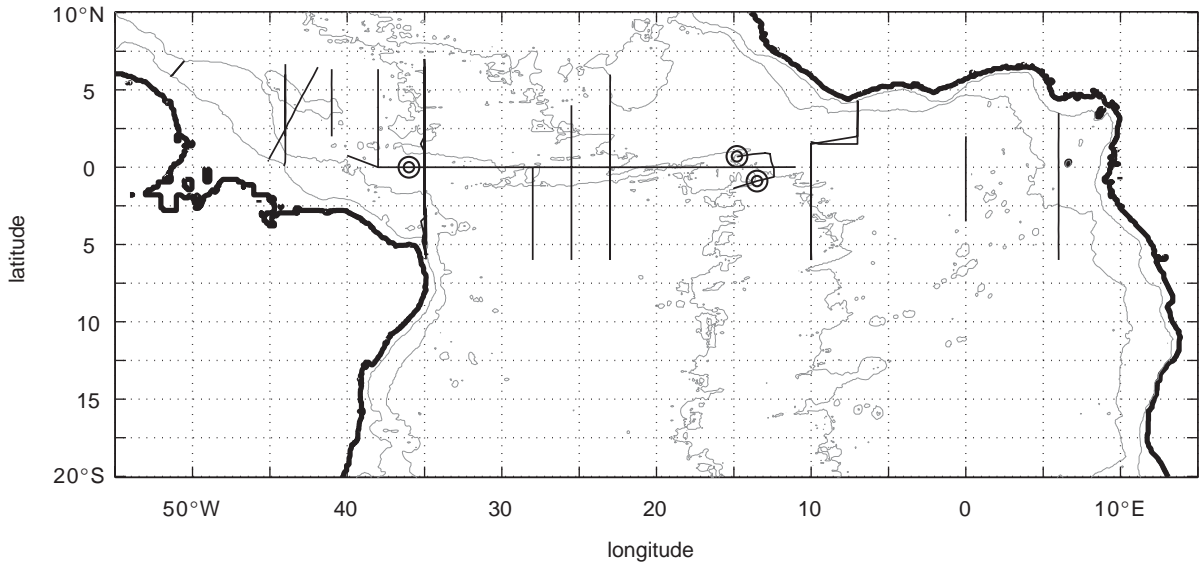


Fig. 1. Tracks of 15 cruises in the tropical Atlantic. The data were obtained in the years 1990–2002. The cruises are: R.V. *Meteor* 14/2, 10/1990; R.V. *Meteor* 16/3, 5–6/1991; R.V. *Meteor* 22/2, 11/1992; Cither 1, R.V. *L'Atlante*, 2/1993; R.V. *Meteor* 27/3, 3–4/1994; Romanche 3, 11/1994; Etambot 2, R.V. *Edwig Link*, 4/1996; R.V. *Seward Johnson*, 7/1997 (three zonal sections, 6°S, equator, 6°N); R.V. *Equalant*, R.V. *La Thalassa*, 7–8/1999; R.V. *Seward Johnson*, 1/2000; R.V. *Meteor* 47/1, 4/2000; R.V. *Equalant*, 7/2000; R.V. *Oceanus* 365/4, 3/2001; R.V. *Ronald H. Brown*, 2/2002. The longitudes of the sections are given in Table 1. Two concentric circles indicate locations of current meter moorings maintained in 1992–1994. See Thierry (2000) for the two eastern moorings and Send et al. (2002) for the western mooring. The isobaths are 1000 and 4000 m.

1996). This result is similar to the error of about  $5 \text{ cm s}^{-1}$  estimated by Fischer and Visbeck (1993). Herein we will assume that the measurement error is about  $5 \text{ cm s}^{-1}$ .

The output from a global model of ocean tides was used to remove the barotropic tides from the velocity measurements. The method was also applied by Schmid et al. (2001), and the limitations were discussed therein. A justification for this correction is also given by Bourlès et al. (2003). The original model is described in detail in Egbert et al. (1994). The largest tidal corrections applied to the zonal velocities presented in this study are  $4 \text{ cm s}^{-1}$  and the mean correction is on the order of  $1 \text{ cm s}^{-1}$ .

Throughout this study a peak-to-peak amplitude on the order of  $20 \text{ cm s}^{-1}$  and a vertical scale of 400–600 m (or alternatively a vertical distance to an adjacent velocity maximum of opposite sign if thick layers with only small velocity changes exist) is used to identify jets.

### 3. Observations

#### 3.1. Zonal and vertical structure

Fig. 2 shows profiles of the zonal velocity obtained on the equator at different longitudes and times. The curvature of the velocity profiles is used to highlight westward anomalies in blue and eastward anomalies in red. In the summer of 1997 (Fig. 2A) five jets of varying thickness are present between the Equatorial Undercurrent and 1600 m. The westward jets are centered near 250, 1100 and 1500 m and the eastward jets are centered near 400 and 1200 m. A transition to another eastward jet is visible below 1500 m. This pattern is interrupted by a thick layer of predominantly westward flow at intermediate depth (about 500–1000 m), where only small changes of the zonal velocity exist. The jet near 250 m is different from the other westward jets. While it has a pronounced westward anomaly in most profiles the total velocity is

Table 1

The longitudes and times of 14 cruises are given

Cruise	Time	Longitudes	Orientation	Instrument type
R.V. <i>Meteor</i> 14/2	10/1990	35°W	Meridional	LADCP
R.V. <i>Meteor</i> 16/3	6/1991	35°W	Meridional	LADCP
R.V. <i>Meteor</i> 22/2	11/1992	35°W	Meridional	LADCP
R.V. <i>L'Atlante, Cither</i> 1	2/1993	35°W	Meridional	Pegasus
R.V. <i>Meteor</i> 27/3	3/1994	35°W	Meridional	LADCP
<i>Romanche</i> 3	11/1994	12.5°W	Meridional	LADCP
R.V. <i>Edwig Link, Etambot</i> 2	4/1996	35°W	Meridional	LADCP
R.V. <i>Seward Johnson</i>	7–8/1997	11–38°W	Zonal	LADCP
R.V. <i>La Thalassa</i> ; R.V. <i>Equalant</i>	7–8/1999	10, 23, 35°W	Meridional	LADCP
R.V. <i>Seward Johnson</i>	1/2000	23, 25.5, 28°W	Meridional	LADCP
R.V. <i>Meteor</i> 47/1	4/2000	35°W	Meridional	LADCP
R.V. <i>Equalant</i>	7/2000	6°E, 0°, 10°W	Meridional	LADCP
R.V. <i>Oceanus</i> 365/4	3/2001	35, 38°W	Meridional	LADCP
R.V. <i>Ronald H. Brown</i>	2/2002	35, 44°W	Meridional	LADCP

The column 'orientation' indicates if zonal or meridional sections were taken, LADCP stands for Lowered Acoustic Doppler Current Profiler.

sometimes eastward, which may be due to the close proximity to the Equatorial Undercurrent.

Three of the five jets (those between 1000 and 1600 m) are clearly visible throughout the 27° long section (the gradual change of the jet depths is discussed below). The two shallowest jets, near 250 and 400 m, can be seen clearly from 38° to 15°W, both in the original and in the smoothed profiles. At 11°W these jets cannot be found in the smoothed profiles, but small anomalies with the correct curvature can be seen in the original profiles. The one corresponding to the 250 m jet is just above 250 m. The other one is at 250 m, i.e. about 150 m shallower than the corresponding jet in the other profiles. The weakness of these anomalies is intriguing because they could be, e.g., indicative for the formation or break-down of the jets.

Two years later (Fig. 2B) the structure differs significantly from the pattern observed in the

summer of 1997. In 1999 the peak-to-peak amplitudes of the zonal velocity and the distances between them are generally more regular than in 1997. In addition, one more jet is observed between the Equatorial Undercurrent and 1600 m in 1999 (a direct comparison of selected profiles obtained at three different longitudes during these two cruises can be found in Figs. 3 and 4). Between 200 and 2100 m the jet structure is very regular, with the peak to peak distance in the range of 200–300 m. Below 2100 m the peak-to-peak amplitudes are smaller and the vertical scale is more variable (not shown).

In January 2000 the flow pattern again deviates from the earlier observations (Fig. 2C; a direct comparison of the 23°W profile with other profiles obtained at the same longitude is shown in Fig. 4A). Two quite broad currents (we do not call them jets because of their large vertical extent) exist between the Equatorial Undercurrent and

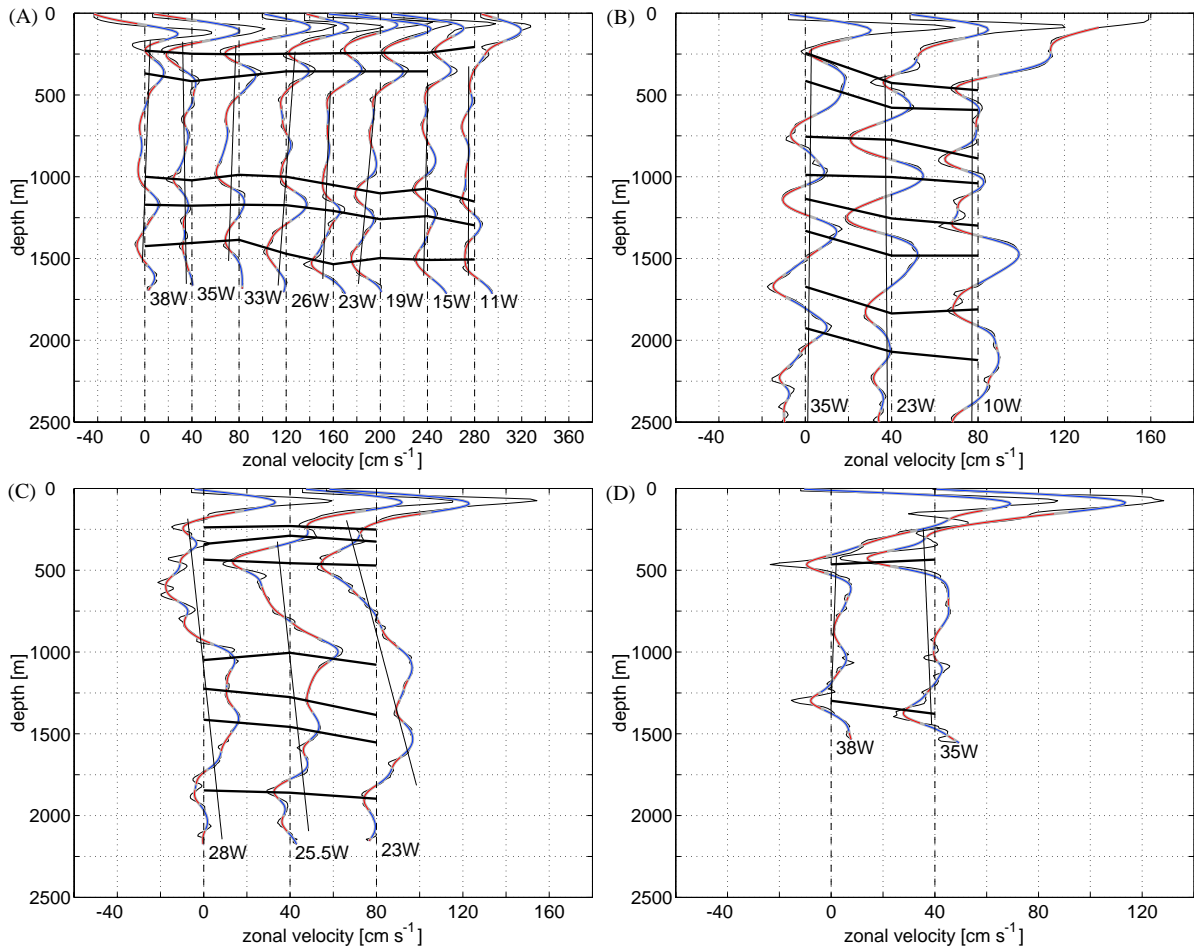


Fig. 2. Profiles of the zonal velocity on the equator from ADCP observations: (A) 19–25 July 1997, (B) 24 July–15 August 1999, (C) 16–24 January 2000, (D) 7–13 March 2001. The profiles are shifted by  $40 \text{ cm s}^{-1}$ , and the zero velocity for each profile is shown as a dashed line. The thin solid line overlaid over each profile indicates the large-scale vertical slope of the flow (the Equatorial Undercurrent has been excluded in the derivation of the slope). The thick gray lines show the velocity profiles after low-pass-filtering with a Butterworth filter using a cut-off length scale of 100 m. The thick black lines mark features that can be traced from profile to profile. Red (blue) segments of the profiles indicate layers with positive (negative) curvatures, which are frequently associated with westward (eastward) jets.

about 1800 m. Superimposed on them are variabilities with smaller scales. The shallower current has a relatively pointed westward peak at about 400 m in the smoothed curves for two of the profiles, whereas the maximum is significantly wider in the smoothed profile at  $28^\circ\text{W}$ . The original profile has nine small peaks with a vertical distance in the range of about 100–200 m that are superimposed on the westward current. Three of these peaks can be associated with small peaks in

the other two profiles (as indicated by black lines). The shallower two of these anomalies are strong enough to remain visible in the smoothed profiles. The deeper eastward current has two relative maxima of the eastward velocity near 1000 and 1500 m that are separated by a minimum near 1300 m. The corresponding vertical scale of 500 m is typical for jets. However, the mean peak-to-peak amplitude is less than  $10 \text{ cm s}^{-1}$  which is significantly smaller than the typical peak-to-peak



amplitude of jets. Below these two currents a weak westward peak exists near 1800 m at 23° and 25.5°W. At 28°W two westward peaks exist around this depth.

In March 2001 the jets are even more poorly developed than in the summer of 1997 (Fig. 2D; for a direct comparison of 35°W profiles see Fig. 3). The most prominent features are two westward jets at 400 and 1300 m. Even so the variability of the primarily eastward velocity is quite weak, this variability is sufficient to classify both currents as jets since they are about 200 m away from a maximum eastward velocity and have a sufficiently large velocity anomaly. A minimum of the eastward flow can be identified at a depth of 800 m at 38°W. A similar minimum exists at 35°W (the velocity actually changes sign, but it is only weakly westward). As in January 2000 several peaks of small peak-to-peak amplitude (less than  $10 \text{ cm s}^{-1}$ ) and small vertical scale (about 100 m) are observed, only this time they are located between 800 and 1200 m.

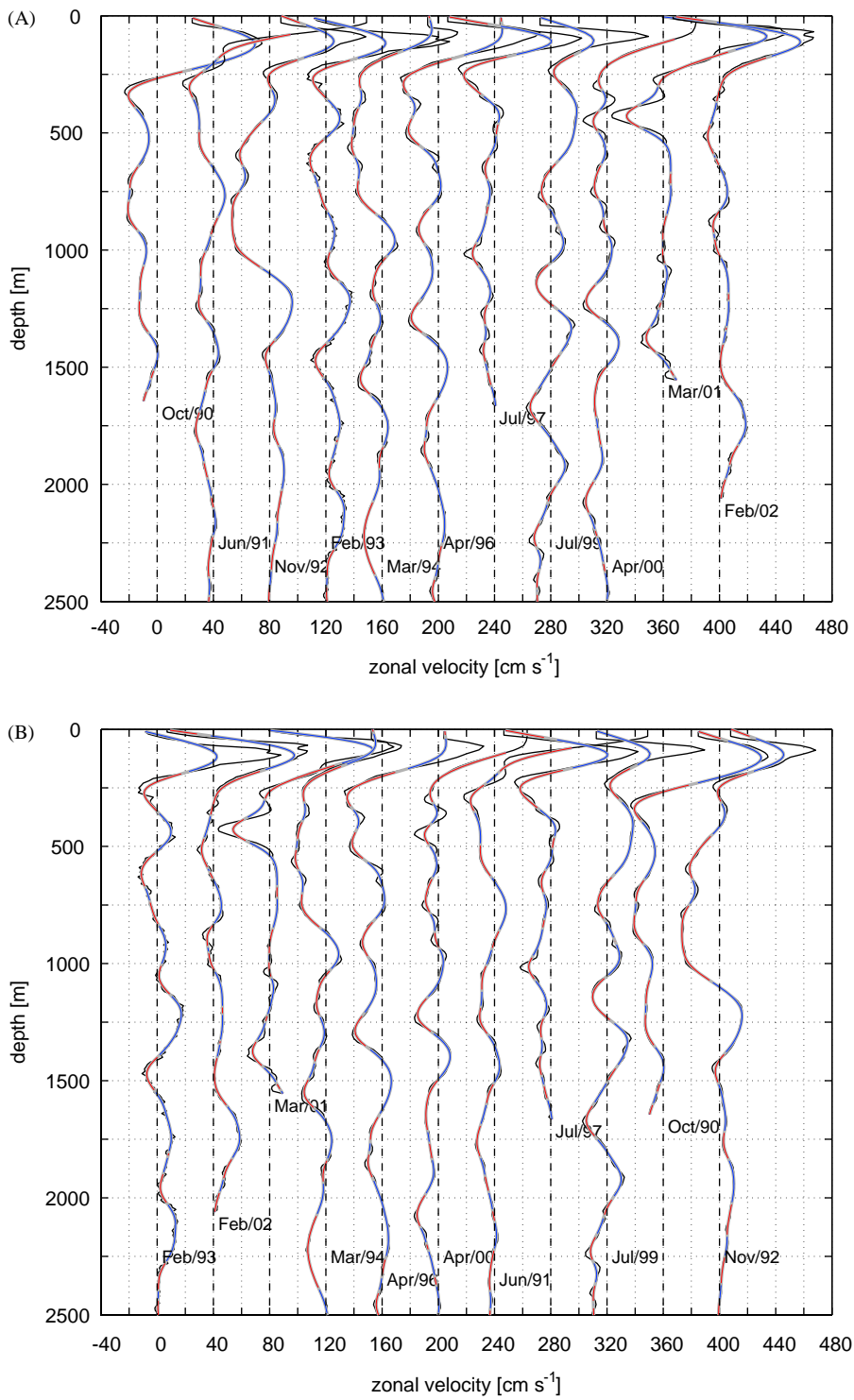
While the velocity profiles obtained in the four surveys differ considerably, they all indicate that changes of the depth of the jets (and some of the other peaks) exist. They are marked with thick black lines in Fig. 2. Throughout this study depth changes are always described in the Eulerian sense. That means the direction of the flow is disregarded. I.e. a jet can shoal from west to east, even if the velocity within the jet is westward. The depth changes listed in Table 2 were estimated by subtracting the two depths of a jet as observed at the eastern and western end of a section, respectively. The resulting value is then divided by the distance in degrees longitude and normalized to be representative for a zonal distance of  $10^\circ$ .

In the summer of 1997 (Fig. 2A) the shallowest two jets shoaled, on average, from west to east by  $5\text{--}8 \text{ m}/10^\circ$  (Table 2). The three jets below 1000 m deepened from west to east by  $44 \pm 13 \text{ m}/10^\circ$  (the given error is the standard deviation). It is noted that the depth changes of these jets depend on the longitude. For the deepest jet, near 1500 m, the depth change is quite large between 33° and 23°W, whereas the depth change outside of this longitude range is rather small and sometimes even of opposite sign (i.e. the jet may be shoaling slightly

towards the east instead of deepening). For the jets near 1000 and 1200 m large depth changes are observed in the longitude range 26–19°W, which fits nicely to the characteristics of the deepest jet near 1500 m. However the two shallower jets also experience large depth changes between 15° and 11°W. In the summer of 1999 no shoaling jet is observed. Instead the eight clearly identifiable jets are deepening from west to east, with a mean depth change of  $62 \pm 21 \text{ m}/10^\circ$ . Consistent with the observations in 1997 most of the deepening occurs between the two western profiles (35° and 23°W). The only exception are the two jets near 750 and 1000 m. In January 2000 six of the seven velocity peaks (as mentioned above none of these peaks has the characteristics typical for jets) are deepening from west to east by  $28\text{--}322 \text{ m}/10^\circ$  (Table 2). Only the poorly defined peak near 300 m shoals slightly. The mean depth change of the six deepening peaks is  $143 \text{ m}/10^\circ$  with a standard deviation of  $124 \text{ m}/10^\circ$ . The relatively large depth changes encountered below 1100 m between 28° and 23°W, the full zonal extent of the survey, is consistent with the observations in the two earlier surveys. In March 2001 the upper westward jet shoals from 38° to 35°W, whereas the lower jet deepens. Both, the shoaling and the deepening, are about twice as large as the maximum shoaling and deepening observed in the same longitude range in the summer of 1997. Because of the limited zonal extents of the January 2000 and the March 2001 surveys the relative large slopes, derived primarily below 1100 m, cannot be seen as representative for basin-scale depth changes. Taken together, these observations suggests that west–east deepening of the jets (and peaks) occurs more frequently than west–east shoaling. In addition, west–east shoaling can primarily be detected at shallower depths (between the Equatorial Undercurrent and 500 m).

### 3.2. Temporal variability

Eleven profiles obtained at 35°W (Fig. 3) are available to analyze the variability of the zonal velocity in the upper 2000 m. In addition, three profiles at 23°W (Fig. 4A) and four profiles in 10–12.5°W (Fig. 4B) were obtained at different times.





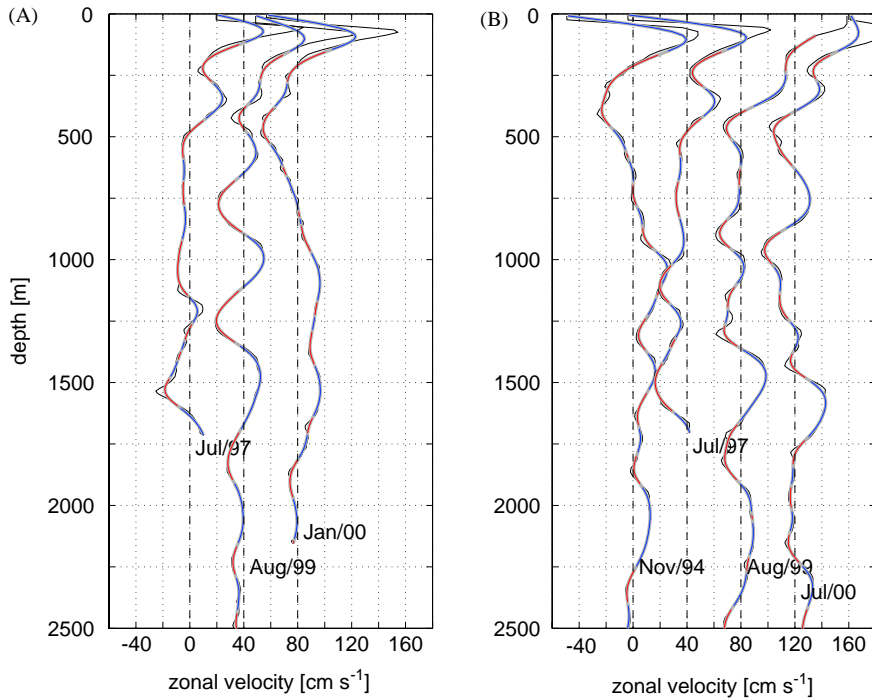


Fig. 4. Profiles of the zonal velocity on the equator at 23°W (A) and in 10–12.5°W (B) from observations with lowered ADCPs. The profiles are shifted by 40 cm s<sup>-1</sup>, and the zero velocity for each profile is shown as a dashed line. The smoothed profiles (thicker gray lines with red and blue overlaid) were obtained by applying a low-pass Butterworth filter with a cut-off length scale of 100 m. Red (blue) segments of the profiles indicate layers with positive (negative) curvatures, which are frequently associated with westward (eastward) jets.

As mentioned in Section 3.1, in some of these profiles the jet structure is strongly developed (e.g., July 1999, April 1996 and February 1993), whereas other profiles have fewer, poorly developed jets with smaller peak-to-peak amplitudes (e.g., July 1997, March 2001). Differences between the jets observed at different times can be characterized as belonging to one of three general categories. At given depths the profiles can have jets pointing in the same direction, they can have opposing jets, or one profile has a zero crossing while the velocity in the other profile is almost at its maximum within a jet. All three configurations can be observed in a comparison of two profiles. A good example are

the profiles obtained at 35°W in February 1993 and March 1994 (Fig. 3A): near 800 m an eastward jet exists in both profiles. Just below 1000 m a westward jet exists in February 1993 (note: this jet can only be identified by the curvature, the total velocity remains weakly eastward in the smoothed profile), whereas a zero crossing was observed in March 1994. Near 2200 m the jets are of opposite sign.

It is also found that none of the profiles obtained within a 1-year long period have matching jets throughout the water column between the Equatorial Undercurrent and the maximum common depth (Fig. 3A). The same is observed for the

Fig. 3. Profiles of the zonal velocity on the equator at 35°W from observations with lowered ADCPs (Pegasus in February 1993): (A) by date, (B) by month. The profiles are shifted by 40 cm s<sup>-1</sup>, and the zero velocity for each profile is shown as a dashed line. The smoothed profiles (thicker gray lines with red and blue overlaid) were obtained by applying a low-pass Butterworth filter with a cut-off length scale of 100 m. Red (blue) segments of the profiles indicate layers with positive (negative) curvatures, which are frequently associated with westward (eastward) jets.

Table 2

Mean depth changes of equatorial deep jets (or velocity peaks if the vertical scale or amplitude differs from those typical for jets), based on the full longitude range of each survey are given

Cruise	Depth (m)	Depth change (m/10°)
<i>Seward Johnson</i> 1997	237	−8
<i>Seward Johnson</i> 1997	374	−5
<i>Seward Johnson</i> 1997	1049	56
<i>Seward Johnson</i> 1997	1213	47
<i>Seward Johnson</i> 1997	1475	30

Mean west–east shoaling:  $-7\text{ m}/10^\circ$ , std.dev.:  $2\text{ m}/10^\circ$

Mean west–east deepening:  $44\text{ m}/10^\circ$ , std.dev.:  $13\text{ m}/10^\circ$

<i>Equalant</i> 1999	380	91
<i>Equalant</i> 1999	528	71
<i>Equalant</i> 1999	806	53
<i>Equalant</i> 1999	1010	20
<i>Equalant</i> 1999	1230	66
<i>Equalant</i> 1999	1432	60
<i>Equalant</i> 1999	1773	56
<i>Equalant</i> 1999	2039	78

Mean west–east deepening:  $62\text{ m}/10^\circ$ , std.dev.:  $21\text{ m}/10^\circ$

<i>Seward Johnson</i> 2000	240	28
<i>Seward Johnson</i> 2000	318	−30
<i>Seward Johnson</i> 2000	454	72
<i>Seward Johnson</i> 2000	1044	58
<i>Seward Johnson</i> 2000	1295	322
<i>Seward Johnson</i> 2000	1475	278
<i>Seward Johnson</i> 2000	1868	102

Mean west–east deepening:  $143\text{ m}/10^\circ$ , std.dev.:  $124\text{ m}/10^\circ$

<i>Oceanus</i> 2001	450	−97
<i>Oceanus</i> 2001	1338	267

For easier comparison all estimates are normalized to be representative for a zonal distance of  $10^\circ$ . Jets (or peaks for the *Seward Johnson* 2000 survey) that are deepening (shoaling) from west to east have positive (negative) values. The means and standard deviations are derived separately for the deepening and shoaling jets (or peaks) observed in the individual surveys.

profiles obtained in the same month or season of different years (Fig. 3B). Some profiles support the presence of an annual cycle, e.g., at 500 m a slight preference for westward flow and westward jets

can be detected in February–April. In July–October the profiles indicate that a preference for eastward flow and eastward jets is possible. However, some profiles do not fit into this pattern, e.g., February 1993 and July 1997. This indicates that other time scales must also play an important role. The temporal variability will be discussed more extensively later, based on estimates of the zonal transport.

### 3.3. Zonal transports

The transports for two water masses will be estimated based on the layer definitions presented in Fig. 5. This figure shows the zonal velocity across three meridional sections between  $28^\circ$  and  $23^\circ\text{W}$  in January 2000 and selected water mass boundaries. The upper and lower boundaries of the AAIW layer are given by the isopycnals  $27.1$  and  $27.4\text{ kg m}^{-3}$  (thick solid lines; Stramma and England, 1999). The core of this water mass is at the depth with the lowest salinity (line with circles). At all three longitudes this layer extends from about 500 to about 1000 m on the equator. The upper limit of the NADW layer is given by the isopycnal marking the depth where the density  $\sigma_2$  is  $36.65\text{ kg m}^{-3}$  (Stramma and England, 1999), near 1200 m. The lower limit of the NADW layer, when defined as the depth of the isopycnal  $\sigma_2 = 37.05\text{ kg m}^{-3}$ , is at about 3900 m (not shown). Before deriving the transports for these two water masses the structure of the zonal velocity will be analyzed.

#### 3.3.1. Antarctic intermediate water (AAIW)

A major difference between the velocity in the AAIW layer of the three sections in Fig. 5 is that at  $28^\circ\text{W}$  in the equatorial band ( $1^\circ\text{S}$ – $1^\circ\text{N}$ ) the whole layer was governed by westward flow (associated with the westward current centered at about 600 m, Fig. 2C), whereas the portion with westward flow was reduced to 75% at  $25.5^\circ\text{W}$  and 50% at  $23^\circ\text{W}$ .<sup>1</sup> Other surveys also reveal significant

<sup>1</sup>Only the southern half of the equatorial band was covered with measurements in the  $28^\circ\text{W}$  section. However, due to the nearly symmetric structure of the equatorial currents and jets, with respect to their vertical extent, it is no problem to speak about the equatorial band in this context.

changes of the flow structure along the equator in this layer. For example, in July 1997 (Fig. 2A) the flow is westward throughout this layer in six out of eight profiles. At 15°W the flow is eastward in the lower half of the layer with a peak velocity of  $7.4 \text{ cm s}^{-1}$ , and at 26°W the flow is eastward in a thin slab with a peak velocity of  $5.3 \text{ cm s}^{-1}$ . Both of these values are larger than the measurement error associated with LADCP data.

At 35°W (Fig. 3) three profiles reveal purely westward flow in the AAIW layer (October 1990, November 1992 and July 1997). Eastward flow throughout most of the AAIW layer is observed in only one profile (March 2001). The other seven 35°W profiles reveal alternating flow within the AAIW layer: In July 1999 westward flow in the AAIW core is surrounded by eastward flow in the upper and lower part of the AAIW layer. The opposite is the case in June 1991 and February 2002, with eastward flow in the core and westward flow above and below the core. In two other profiles (February 1993, March 1994) the upper part of the AAIW layer has westward velocities, whereas eastward velocities are observed in the lower part of this layer.

The components and the total zonal transports across four longitudes, nominally between 1°S and 1°N,<sup>2</sup> are given in Table 3. In addition, Fig. 6 shows the cumulative transports at 35°W (integrated from top to bottom). In the AAIW layer (500–1000 m) at 35°W the largest eastward transport is  $7.8 \text{ Sv}$  ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) and the largest westward transport is  $-24.3 \text{ Sv}$ . The two sections without any eastward transport have large total westward transports and were obtained in the same season (October/November). The largest eastward and smallest westward transports were observed in two summer surveys (June 1991/July 1999), which results in a total eastward transport. All surveys taken together support earlier studies

which concluded that an annual cycle exists (see Section 4), and that the annual mean of the total transport of AAIW may be westward. The latter can be inferred since most months are governed by westward flow and the westward transports can be significantly larger than the largest eastward transports. Interannual variability must also play a role: while no zonal transports could be estimated for the zonal section in July 1997 the flow in the AAIW layer is predominantly westward at most longitudes. Therefore, it can be deduced that the July 1997 transport in the AAIW layer would be in the opposite direction than the transport in July 1999.

As already indicated by the depth changes of the jets and peaks highlighted in Fig. 2 a significant longitude dependence of the transports exists. Two examples, January 2000 and July 1999, are given in Table 3. In the January survey the total transport changes from westward in the two western sections to small and eastward in the easternmost section. In the July survey the total transport changes from eastward at 35°W to a negligible transport at 23°W.

### 3.3.2. North Atlantic deep water (NADW)

As in the AAIW layer alternating zonal velocities are common in the NADW layer. The likelihood for this to occur is even larger than in the AAIW layer, because the NADW layer occupies a significantly larger chunk of the water column.

When comparing the transport estimates it has to be kept in mind that some of the profiles within 1° of the equator do not reach the lower limit of the NADW layer at 3900 m (sections where this is the case are indicated in Table 3). The meridional integration of the transports shown in Fig. 6 was terminated at the depth where more than two profiles did not have velocity data. For the vertically integrated transports this lack of data will lead to an error of the westward and/or eastward transport components. The transports around this depth are smaller than those above and, typically, only about 10% of the data in one profile are not available. From Fig. 6 it can be seen that the error for the components of the transports across a section caused by this lack of data is

<sup>2</sup>Some of the sections do not cover the full latitude range, as indicated in the last column of Table 3. In this context the meridional asymmetry of the currents and jets due to the often off-equatorial location of the velocity maximum cannot be ignored. Therefore, only sections that have observations on both sides of the equator were used for the transport estimates. For the sections that do not cover the equatorial band fully the transports are likely to be underestimated.

mostly small. In April 1996 the total transport is basically negligible, and the shape of the curve indicates that this is unlikely to change if the full depth profiles had been measured. Similarly, the April 2000 transport does not change much near 4000 m. For the July 1999 survey it can be deduced

that the actual westward transport is smaller than the one given in Table 3.

The transports of NADW range from  $-24.9$  to  $22.7$  Sv when eastward and westward flow is treated separately. Even though only relatively few transport estimates are available the observations indicate that an annual cycle may exist. The smallest westward transports of NADW ( $-10.0$  Sv or less) are found in the winter (November, February, Table 3). The smallest eastward transports are observed in April and in June/July. The latter are also the months with the largest westward transports. This results in total transports that have a westward maximum in June/July (about  $-10$  Sv) and an eastward maximum in November (about  $18$  Sv). The transition between these two distinct seasons occurs in spring (March/April, with an eastward transport that is less than  $1$  Sv, with the exception of  $-5.5$  Sv in April 2000) and in the late summer or early fall (not enough data are available to be more precise). It is noteworthy, that the eastward maximum of the total NADW transport coincides with the westward maximum of the total AAIW transport, and that the largest total westward NADW transport was derived for a month when the total AAIW transport is eastward. Overall the annual cycle of the NADW transport is anti-correlated with the AAIW transport (the correlation coefficient is  $-0.86$ ).

#### 4. Discussion and conclusions

Vertical scales ranging from 100 to 1000 m are observed in equatorial profiles of the zonal

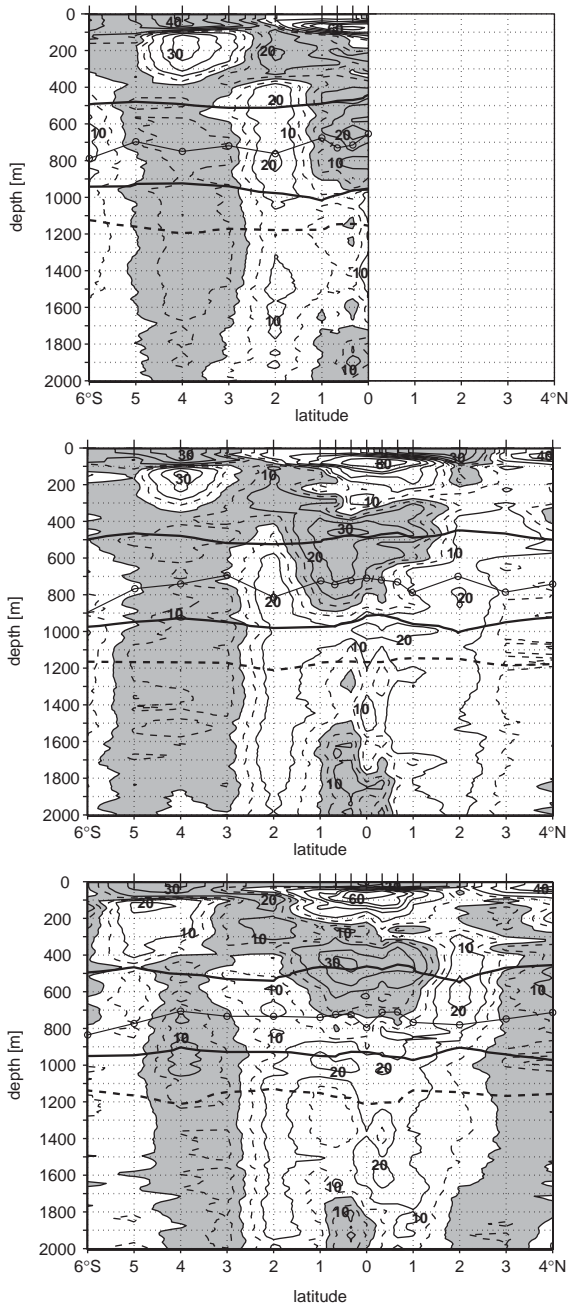


Fig. 5. Sections of the zonal velocity from lowered ADCP measurements at  $28^{\circ}\text{W}$  (top, 22–23 January 2000)  $25.5^{\circ}\text{W}$  (middle, 14–16 January 2000) and  $23^{\circ}\text{W}$  (bottom, 17–20 January 2000). The  $5\text{ cm s}^{-1}$  isoline is dashed. The two thick lines show the depths where the density ( $\sigma_{\theta}$ ) is  $27.1$  and  $27.4\text{ kg m}^{-3}$  to indicate the upper and lower boundary of the Antarctic Intermediate Water (AAIW) layer. The thin line with circles marks the salinity minimum of the AAIW. The thick dashed line shows the depth where the density ( $\sigma_2$ ) is  $36.65\text{ kg m}^{-3}$  to indicate the upper boundary of the North Atlantic Deep Water layer. Tic marks at the top of each panel indicate the profile locations. The equatorial profiles for these sections are shown in Fig. 2C.

Table 3

Zonal transports within 1° of the equator in the AAIW layer (500–1000 m) and in the NADW layer (1200–3900 m)

Date	AAIW transport (Sv)			NADW transport (Sv)			Latitude range
	Total	Eastward	Westward	Total	Eastward	Westward	
35° W							
02/1993	−2.1	2.1	−4.2	8.7	18.7	−10.0	0.8°S–0.6°N
02/2002	−1.5	1.9	−3.4	—	—	—	1.0°S–0.4°N
03/1994	−6.2	2.9	−9.1	0.8	19.0	−18.2	1.1°S–0.6°N
03/2001	−5.6	0.7	−6.3	—	—	—	0.7°S–1.0°N
04/1996	−2.9	2.8	−5.7	0.3 <sup>1</sup>	16.4	−16.1	1.0°S–1.0°N
04/2000	−5.3	0.7	−6.0	−5.5 <sup>1</sup>	12.5	−18.0	1.0°S–1.0°N
06/1991	4.0	6.2	−2.2	−10.2	11.4	−21.6	1.1°S–0.7°N
07/1999	6.7	7.8	−1.1	−9.7 <sup>1</sup>	15.2	−24.9	1.0°S–1.0°N
10/1990	−13.4	0.0	−13.4	—	—	—	0.4°S–0.5°N
11/1992	−24.3	0.0	−24.3	17.9	22.7	−4.8	1.1°S–0.6°N
28° W							
01/2000	−6.1	0.4	−6.5	—	—	—	1.0°S–0.0°N
25.5° W							
01/2000	−4.2	5.5	−9.7	—	—	—	1.0°S–1.0°N
23° W							
07/1999	0.1	5.6	−5.5	−7.6 <sup>4</sup>	12.9	−20.5	1.0°S–1.0°N
01/2000	0.9	6.3	−5.4	—	—	—	1.0°S–1.0°N

For some sections no NADW transports could be estimated because too many profiles did not extend to 3900 m. The transports (negative → westward) were derived from the LADCP data collected along cross-equatorial sections. 1 Sv is  $10^6 \text{ m}^3 \text{ s}^{-1}$ . The superscript numbers in the column of total NADW transports give the number of profiles with a maximum depth that is less than 3900 m. One of them ends at 3192 m, the others end between 3480 and 3560 m.

velocity collected in the tropical Atlantic. Among them are the scales of 400–600 m which are typical for the equatorial jets in the upper 2500 m in the Atlantic (Gouriou et al., 1999; Johnson and Zhang, 2003). Scales in the upper range (about 1000 m) indicate the absence of jets in a part of the water column. Scales in the lower range (about 100 m) can be present within the jets or in a jet-free depth interval. In both cases the peak-to-peak amplitudes of the zonal velocity are smaller than the about  $20 \text{ cm s}^{-1}$  typical for the jets.

Observations at eight longitudes reveal that the jets in the Atlantic can have a zonal extent of at least  $27^\circ$  (3000 km, in the summer of 1997, Fig. 2A). Previously, Gouriou et al. (2001) came to the conclusion that the zonal extent may be  $25^\circ$  (based on observations at only three longitudes). The larger zonal resolution increases the confidence

that the derived larger value for the length scale is correct. In this and other sections a preference for west–east deepening of the jets and velocity peaks is apparent (Fig. 2). The depth change is mostly about  $50 \text{ m}/10^\circ$  in the zonal averages, and can be as large as  $320 \text{ m}/10^\circ$  (Table 2). The higher values are derived for surveys that only cover short distances ( $5^\circ$  or less). There are strong indications that the west to east deepening of the jets and velocity peaks may be largely concentrated between  $35^\circ$  and  $23^\circ \text{W}$ . This is the reason why some of the surveys of limited zonal extent can have large mean depth changes.

The large zonal extent of the jets and the changes of their depth along the equator are in general agreement with long vertically propagating linear waves (McCreary, 1984). For example, McCreary (1984) derived depth changes of about

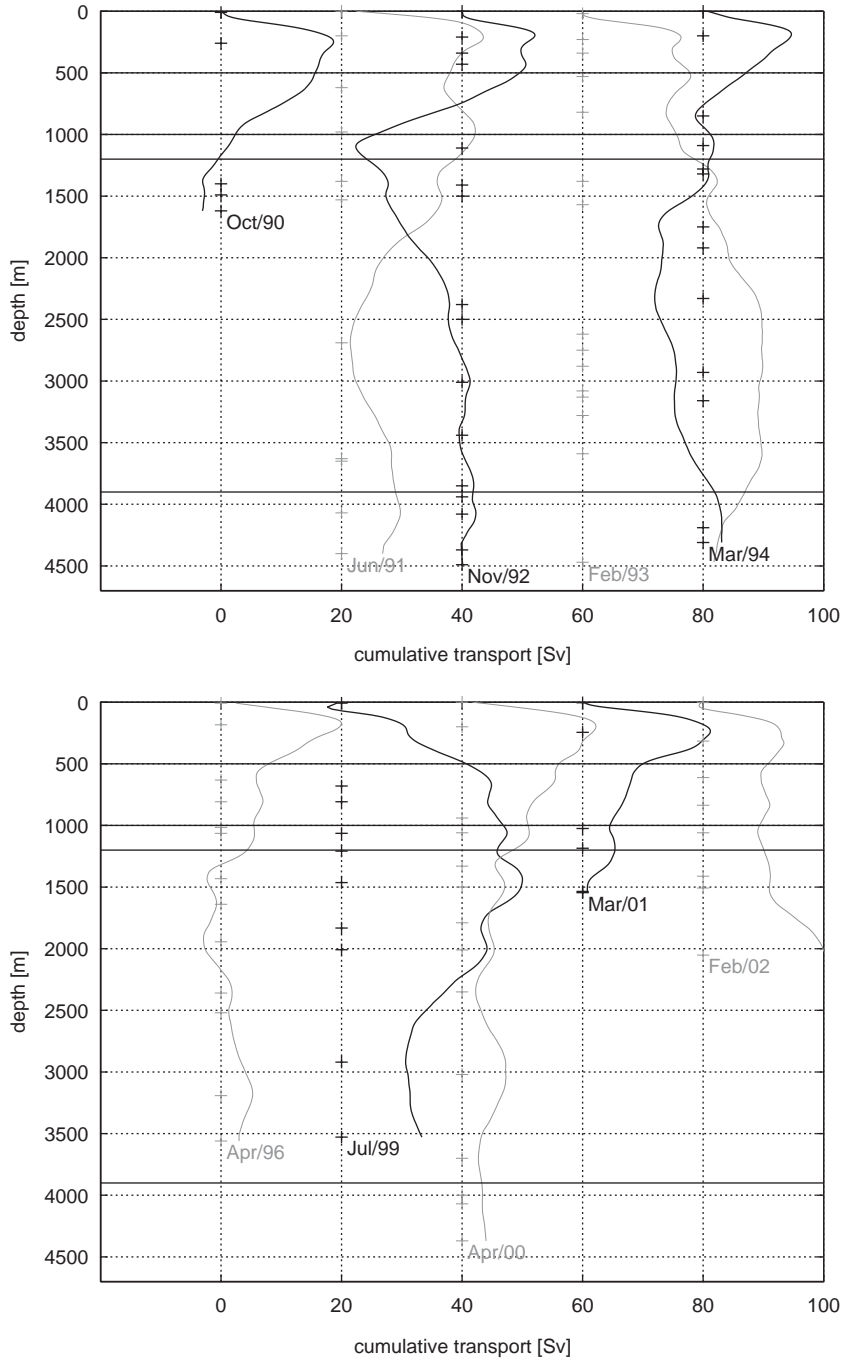


Fig. 6. Profiles of the zonal cumulative transport (integrated from top to bottom) at 35°W from observations with lowered ADCPs (Pegasus in February 1993). The profiles are shifted by 20 Sv ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ). The crosses indicate the depths where the direction of the transport changes. The straight lines indicate the boundaries chosen for the Antarctic Intermediate Water (500–1000 m) and North Atlantic Deep Water (1200–3900 m) layer.



140 m/10° and 380 m/10° for the first and third meridional mode of equatorially trapped Rossby waves (his Fig. 6A), respectively, and about 50 m/10° for Kelvin waves. These values are within the range of values derived here (Table 2). It has to be cautioned, however, that slopes as large as those derived for Rossby waves are only observed within a limited longitude range. The longitude dependence of the slopes indicates that the simple linear wave theory (aside from other limitations discussed below) is not sufficient to describe the characteristics of the jets. A possible reason for this may be topographic effects. On the equator the Mid-Atlantic Ridge covers two broad regions that are connected by a narrow ridge sloping from the Northeast to the Southwest (Fig. 1). The western region extends from 30° to 22°W, with a water depth of 2200–3400 m. The eastern region extends from 19° to 13°W, with a water depth of, mostly, 2800–3200 m. In both longitude bands the smaller water depths are found within 0.5–1° of the equator. The shallowest water depth, 1600 m, is found near 0.5°N, 17°W. The western region is within the longitude range where the depth change of the jets and peaks is larger, and closer to the values predicted by the linear wave theory (e.g., for three jets observed in July 1997, Fig. 2A, for some of the peaks observed in January 2000 and for one jet observed in March 2001). In the eastern region, only two jets reveal a relatively large depth change in the same section (near 13°W). Consistent with the differences between the observations and the linear wave theory found here, several studies (e.g., Pedlosky, 2003; Muench and Kunze, 2000; Hua et al., 1997; Muench et al., 1994) have suggested that the linear wave theory alone is not sufficient to explain the jets. However, Rossby and Kelvin waves are likely to play a significant role.

The observations show that the profiles of the zonal velocity can have opposing jets at some depths and jets pointing in the same direction at other depths (Figs. 3 and 4). Such cases occur for profiles obtained in the same month of different years as well as for profiles obtained within 4 months of each other. In addition, during some surveys the jet structure is very strong and regular, whereas they are only poorly developed in other

surveys. Firing (1987) reported that jets observed in a 16-month time series can disappear and reappear again at some depths, while some of those at other depths remain visible over most of the period of observation. Firing (1987) also showed that many of the jets shoaled 50 m over the time period, and that a vertical displacement occurred between 1980 and 1982/1983. According to Firing (1987) the latter could be, e.g., due to a local deepening of about 130 m, or a local shoaling of 220 m. Alternatively, it could be that the jets form, grow and decay in place, or that the jets change depth by decaying at one depth and forming at another depth (Firing, 1987). The latter two possibilities could be the explanation for the profiles with poorly developed jets in part of the water column. Taken together these observations indicate that the temporal variability of the jets is complicated.

The jets and velocity peaks have a significant impact on estimates of the transports of the Antarctic Intermediate Water (AAIW) and the North Atlantic Deep Water (NADW), since they are dynamically independent of the vertical water mass distribution, and since their vertical extent is smaller than the thickness of the layers occupied by the water masses. For this reason both eastward and westward transports can occur at the same time within a water mass (Table 3).

In the AAIW layer the total transport is in the range of –24.3 to 6.7 Sv. During most surveys the transport of AAIW is westward. Only in June 1991 and July 1999 the total transports were eastward. In July 1997 no transport could be estimated, but the flow structure indicates that the transport would be westward. This indicates that an annual cycle may exist, and that interannual variability may be important. Current meter records presented in various studies (e.g., Weisberg and Horgan, 1981; Eriksen, 1985; Thierry, 2000; Send et al., 2002) allow the identification of important time scales associated with the variability of the zonal flow. The mooring records for the zonal velocity presented by Eriksen (1985) indicate that a period of 1–2 months with a peak-to-peak amplitude of about 10 cm s<sup>-1</sup> can exist at depths of 500–600 m (primarily during the second half of the year, i.e. at the time when tropical instability

waves are strong). The importance of this observation is limited since Eriksen's moorings were in the Pacific. The other mooring records (Weisberg and Horigan, 1981; Thierry, 2000; Send et al., 2002) were located in the Atlantic and they do not show such a mesoscale variability in the zonal velocity. The mooring records available to Thierry (2000) have some higher-periodic variability, but the largest peak-to-peak amplitudes of about  $5 \text{ cm s}^{-1}$  occur in the first half of the year. It has to be cautioned here, that only Weisberg and Horigan had moorings in the depth range above 1000 m. If variability with mesoscale periods is present in the mooring records it is characterized by an amplitude that is smaller than the amplitude associated with the jets. Therefore mesoscale variability does not play a significant role for the discrepancies between the sections obtained during the same months of different years. It seems more likely that these discrepancies are due to inter-annual variabilities. The transport estimates from these surveys and the knowledge about the mesoscale variability suggest that an annual cycle exists. These results are consistent with velocity observations published earlier. For example, quasi-Lagrangian observations from profiling floats showed that the westward flow in July 1997 continued until November 1997 in the lower part of the AAIW layer (Molinari et al., 1999; Schmid et al., 2001). The flow then reversed to eastward until the floats left the equatorial band in January 1998. Ponte (1989) showed a Pegasus profile obtained in January 1989 at  $30^\circ\text{W}$  with eastward and westward jets governing the flow in the AAIW layer, resulting in a mean flow towards the east. This is consistent with the quasi-Lagrangian float observations obtained 9 years later, especially since the lower part of the AAIW layer has eastward velocities (as found here for January 2000). Boebel et al. (1999) combined earlier observations (from 1974 to 1994, various instruments; Weisberg et al., 1980; Weisberg, 1980; Richardson and Schmitz, 1993; Schott et al., 1998) with newer observations (from 1994 to 1995, RAFOS floats). Most of these observations are from the western Atlantic ( $22\text{--}45^\circ\text{W}$ ). Boebel et al. (1999) concluded that the water in the AAIW core flows eastward in April to July and westward in

August to November, whereas it can be transported either way in the other months. While direct comparisons of their and our results are difficult, since they were derived for different depth ranges, all observations suggest that eastward transports of AAIW occur in early summer (and sometimes also in spring), whereas westward transports prevail for the rest of the year.

Because of the large temporal variability, it was not attempted to use the transports in Table 3 to estimate the annual mean of the transport of AAIW along the equator. But it is noted that the prevalence and the larger magnitude of westward transports indicates that the mean transport may be westward. An estimate of the mean zonal transport across  $35^\circ\text{W}$  between 700 and 900 m was obtained by Schott et al. (2003). They derived a westward transport of 10 Sv on the basis of twelve LADCP sections and an assumption that an annual and a semi-annual cycle dominate the variability of the transport.

In contrast to the evidence from the velocity observations, studies of water mass distributions have indicated that the mean transport of AAIW along the equator may be weaker than the velocity-based results suggest. Maps of the salinity on two adjacent isopycnals in the AAIW layer ( $27.2$  and  $27.3\sigma_\theta$ ) give no indications for either eastward or westward spreading of this water mass along the equator (Tsuchiya et al., 1994; Suga and Talley, 1995). Warner and Weiss (1992) derived a map of the distributions chlorofluoromethane (CFM) concentration on the isopycnal  $27.2\sigma_\theta$  (their Fig. 11) and analyzed two CFM sections along the prime meridian. Their map shows a broad region of higher concentrations near the equator. Within this region the concentration decreases towards the east, which they interpreted as a sign for eastward flow. The highest concentrations are found south of the equator at the latitude of the Southern Intermediate Counter-current (SICC, see Stramma and Schott (1999) for a review and Schmid et al. (2003) for additional details). The map by Warner and Weiss (1992) indicates that the distance between the maximum in the tongue and the equator decreases from west to east. At the prime meridian the maximum is wide enough to include the equator. This may be

due to horizontal mixing of equatorial water (lower CFM concentration) with the water in the eastward SICC (higher CFM concentration) that can cause a faster reduction of the CFM concentration in the SICC and slower reduction of the CFM concentration on the equator. Signs for water exchanges consistent with this process were detected in RAFOS trajectories by Boebel et al. (1999). The distribution observed by Warner and Weiss (1992) can exist during periods of eastward or westward equatorial flow if the mixing is strong enough, i.e., the change of CFM concentrations along the equator are not sufficient to determine direction of the mean equatorial flow.

Model results by Jochum and Malanotte-Rizzoli (2003) also suggest that the zonal AAIW transport is governed by an annual cycle and that its mean transport along the equator is weak. It has to be noted that the vertical resolution of their model (20 levels between 100 and 3000 m) is not sufficient to simulate the jets. This may be why the zonal velocities in their model between the Equatorial Undercurrent and 1500 m have peaks of about  $10 \text{ cm s}^{-1}$ , whereas the observed velocity peaks can have  $20\text{--}30 \text{ cm s}^{-1}$ . If the jets are persistent features, as suggested by some observations (see Section 1), then their absence in the model can prevent the derivation of realistic mean transports on a year-by-year basis. In this case only long-term means of the transport may be realistic, but only if the transports due to the jets cancel out.

In the NADW layer the total transports range from  $-10.2$  to  $17.9 \text{ Sv}$  (Table 3). The observations indicate that an annual cycle may exist at  $35^\circ\text{W}$ , with westward total transports in the summer and eastward total transports in the winter. The uncertainty of the inference of an annual cycle for the transport of NADW is larger than for the transport of AAIW because fewer sections were available. However, it is noted that the total transports in the AAIW layer are often of opposite sign when compared with the total transport of the NADW layer (the correlation coefficient is  $-0.86$ ) which increases our confidence in the presence of an annual cycle of the transport in the NADW layer. Also, this annual cycle is consistent with independent observations in the central equatorial Atlantic: Thierry (2000) reported the observation

of annual cycles in the Romanche Fracture Zone (at  $0.7^\circ\text{N}$ ,  $14.8^\circ\text{W}$ , Fig. 1) and the Chain Fracture Zone (at  $0.9^\circ\text{S}$ ,  $13.5^\circ\text{W}$ ) using current meter records obtained at depths of 1700 and 2000 m between November 1992 and October 1994. The peak-to-peak amplitude associated with this signal was  $10\text{--}20 \text{ cm s}^{-1}$ , and the flow was mostly eastward in fall/winter and mostly westward in spring/summer, which is consistent with our result.

A similar annual cycle was also detected in the zonal velocity fields of a model experiment at depth of 1875 m between  $5^\circ$  and  $35^\circ\text{W}$  on the equator (Böning and Schott, 1993, their Fig. 7). The vertical grid spacing at this depth is 250 m, which is not sufficient to fully resolve the jets. Therefore the annual cycle in the model cannot be explained by the jets. Based on a time series from current meters at four depths (1300, 1500, 1700 and 1900 m), Send et al. (2002) reported that no seasonal reversals of the zonal flow occurred on the equator at  $36^\circ\text{W}$  (Fig. 1) during a 20-month period (October 1992–May 1994). However, the current meter at 1500 m recorded variations of the zonal velocity with a peak-to-peak amplitude of  $10\text{--}30 \text{ cm s}^{-1}$  and periods of 4–6 months. This range of periods is comparable to results presented by Weisberg and Horgan (1981) who reported periods of 4–8 months in the Gulf of Guinea (at 869 and 1284 m, their record at 1822 m is too short to detect such periods). Throughout the first 460 days of the current meter record presented by Send et al. (2002) the amplitude of these variations was not large enough to cause reversals of the zonal flow.

The discrepancy between the current meter records available to Thierry (2000; reversals of the flow at an annual period) and Send et al. (2002; no reversals of the flow at an annual period) is very interesting, since they were obtained for nearly the same time periods. It seems possible that the variability in the central Atlantic may have a different characteristic than the variability in the western Atlantic. This would mean that the similarity of the annual cycle of the NADW transports derived from the synoptic sections in the western Atlantic and the current meter records in the central Atlantic is only a coincidence that may, in fact, be caused by interannual variabilities

in the western Atlantic (as suggested by Send et al., 2002). However, this does not explain the quasi-semiannual period found in the 1500 m mooring.

No conclusive theory has been presented that explains the differences between the current meter records obtained at these different locations. Several processes could cause such differences: (1) Effects due to the close proximity of the western boundary with its own dynamics may have a large enough influence to overshadow an annual cycle at 36°W. This could be caused by an off-shore recirculation of the Deep Western Boundary Current. McCartney (1993) summarized and confirmed earlier studies that indicate that such a recirculation exists. Rhein et al. (1995) analyzed tracers and transports which also support a recirculation of the Deep Western Boundary Current. They found, for example, that the transport of the shallow upper NADW in the Deep Western Boundary Current increased from 4 to 11 Sv between 47° and 35°W. Based on their schematic, the recirculation is likely to have a significant impact on the velocities observed on the equator at 36°W. (2) The Mid-Atlantic Ridge may prevent the penetration of an annual wave signal from the eastern into the western basin. Barnier (1988) performed a model study in the North Atlantic that indicates that the Mid-Atlantic Ridge can block planetary waves at mid-latitudes. A result of this is a generation of planetary waves with different properties on the western side of the ridge. It has to be cautioned, however, that the impact of the Mid-Atlantic Ridge on equatorially trapped waves may be different, because their propagation has a large vertical component. A study of velocities from a general circulation model by Li and Chang (1999) shows that an annual cycle is present on the equator on both sides of the Mid-Atlantic Ridge, but there are signs that the ridge has an impact on the phase of the waves. As in the model fields used by Böning and Schott (1993) the vertical grid spacing is too large to resolve the jets (it is about 500 m at the depth of the NADW).

It is not clear at this time if the observations of an annual cycle are due to a variability of the jets, or due to a variability of currents with a larger vertical extent than the jets. The model studies by

Böning and Schott (1993), Li and Chang (1999), and Jochum and Malanotte-Rizzoli (2003) are in favor of the latter, but their validity to address this question is limited because of the insufficient vertical resolution. The anti-correlation of the AAIW and NADW transports also supports the concept that currents of larger vertical scales may be the primary cause for the annual cycle. This is consistent with results from some earlier studies which reported quite stable jet structures over extended periods of time (one year or even longer; e.g., Muench et al., 1994; Send et al., 2002), which indicates that interannual rather than shorter-periodic variability might give rise to the differences between the jet structures observed in profiles from the same season and longitude (Figs. 3 and 4).

An experiment to address the open questions about the jets would have to involve current meter moorings at several longitudes along the equator (every 5–10°, with measurements on both sides of the Mid-Atlantic Ridge) with a vertical resolution of at least 50 m. Numerical models would need a similar or higher vertical resolution to be able to simulate the jets.

## Acknowledgments

We want to thank Christopher Meinen and Rick Lumpkin for helpful suggestions and Friedrich A. Schott for making the data from several cruises available to us. We are also grateful for the software to estimate tidal currents which was provided by Eric Firing and colleagues. This work is partially sponsored by NOAA's Office of Global Programs.

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