Comments on "Mesoscale Variability in the Atlantic Ocean from Geosat Altimetry and WOCE High-Resolution Numerical Modeling"

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1. Introduction

In a recent paper, Stammer and Böning (1992), hereafter noted as SB, made a systematic analysis of wavenumber spectra and spatial scales of mesoscale variability deduced from Geosat altimetric data in the Atlantic. They compare their results with those deduced from the WOCE Community Model Experiment (CME). Geosat data had already been analyzed by Le Traon et al. (1990, hereafter noted LRB) in the North Atlantic. The two data analyses, naturally, lead to similar results. However, their interpretation by SB and LRB are very different. The purpose of this comment is to discuss these differences and clarify the interpretation of wavenumber spectra and spatial scales of altimetric mesoscale variability.

2. Do altimetric wavenumber spectra follow a k^{-5} law?

Stammer and Böning claim that Geosat wavenumber spectra of sea level anomaly (SLA) (or dynamic height anomaly) follow a k^{-5} law over almost the entire Atlantic Ocean except for certain tropical and subtropical areas. They suggest that departures from the k^{-5} law are due to measurement errors. However, the very same Geosat spectra seem to contradict these conclusions. In most areas (not only in the tropics and subtropics), a simple least-squares fit shows that the spectra tend to follow k^{-2}–k^{-3} laws. The steeper slopes observed in high energy areas are also better fitted with a k^{-4} law than with k^{-5}. Figure 1 taken from LRB is a good example of what spectra look like in the North Atlantic. Similar results are found in SB (e.g., see their Fig. 6). Figure 1 shows clearly that altimetric spectra do not follow a uniform k^{-5} law, a finding that agrees with Fu (1983) and Fu and Zlotnicki (1989).

Are the departures from the k^{-5} law due to measurement errors (instrumental noise or geophysical errors) as SB claim? Instrumentation noise is a white noise of about 150 cm^2/cycle/km for Geosat and does not affect spectra for wavelengths longer than about 50–100 km (depending on latitude). Numerous studies have shown that Geosat data are not overcontaminated by geophysical errors (e.g., Fu 1983; Bisagni et al. 1989; Zlotnicki et al. 1989; Jourdan et al. 1990; Le Traon et al. 1990; Monaldo 1990). Spectral analysis of these errors (e.g., Fu 1983; Jourdan 1990) shows indeed that for wavelengths shorter than 1000 km, the oceanic signal is only slightly affected. This means that mesoscale spectra are only slightly affected. Comparison with in situ measurements has also shown that Geosat data are reliable even in the low-energy areas where SB questioned the quality of Geosat data. For example, Stammer et al. (1991) have found very good correlation of Geosat data with hydrographic data in the Iberian Basin. De Mey et al. (1993) reached the same conclusion in the northeast Atlantic. One can thus conclude that wavenumber spectra of sea level anomaly deduced from altimetry do not follow a uniform k^{-5} law and that this is not due to measurement errors.

3. Interpretation of altimetric wavenumber spectra

Le Traon et al. interpret wavenumber spectra for energetic areas similarly to SB. In these areas, the dominant mode of generation of eddy energy is the barotropic and baroclinic instabilities of the mean currents. Wavenumber spectrum slopes, although weaker, are close to the values given by the theory of quasigeostrophic turbulence (Charney 1971). Simulations of quasigeostrophic turbulence (Hua and Haidvogel 1986) have also shown that spectral peaks occur generally at scales of roughly twice the first internal Rossby radius. This is slightly lower than the barotropic β arrest scale k_β (Rhines 1977), that is, where the dispersion of Rossby waves begins to dominate the ocean signal. LRB have shown that the shapes of altimetric spectra are similar to those obtained by numerical simulation, in terms both of the spectral peak and of the behavior beyond the break where the energy level decreases very fast. In lower-energy regions, one of the possible generation mechanisms for eddy and low-frequency mo-

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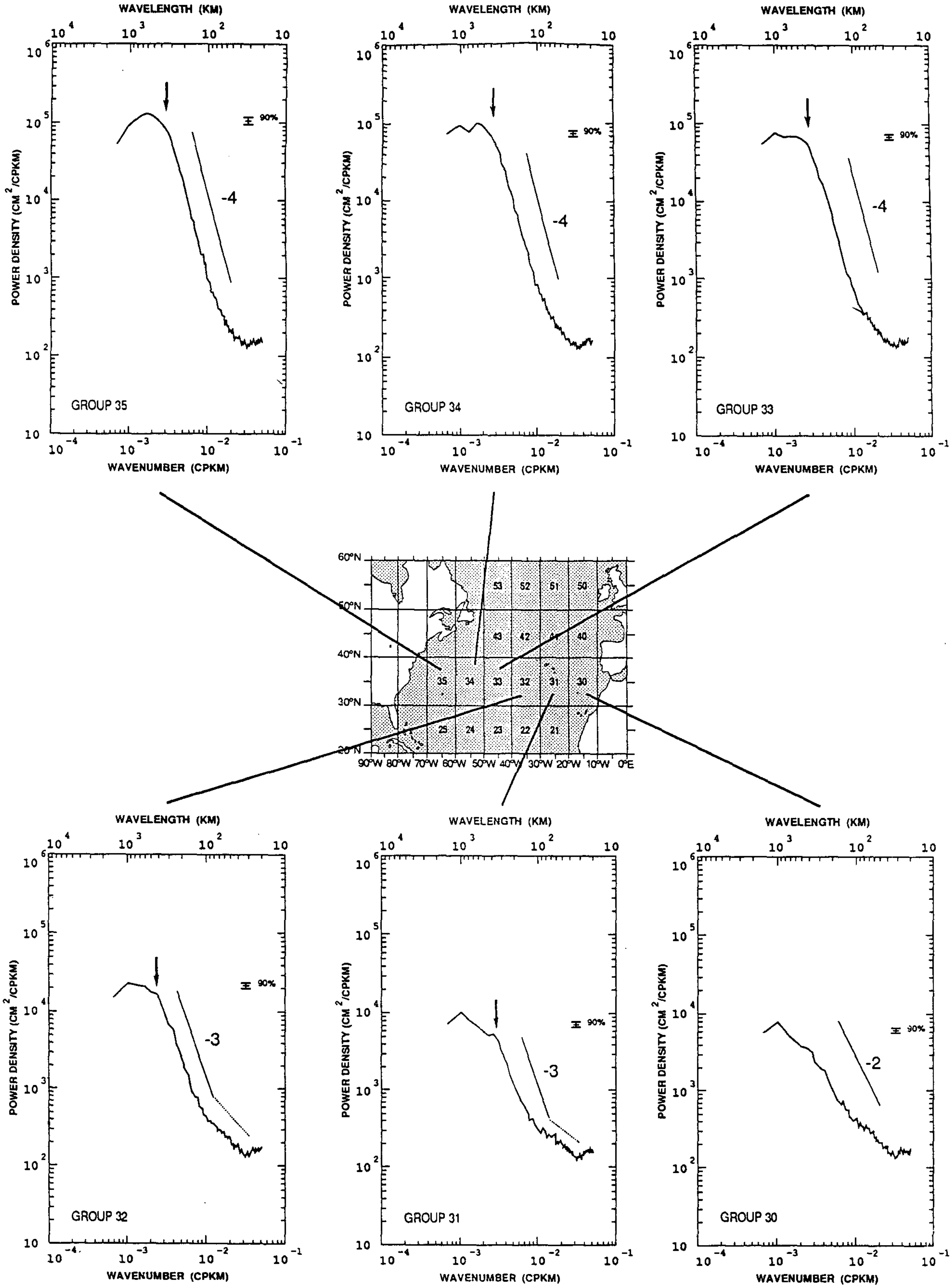
WAVELENGTH (KM)
POWER DENSITY (CM^2/CPKM)
WAVENUMBER (CPKM)
90%
GROUP 35
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tions is forcing by fluctuating winds (Frankignoul and Müller 1979; Müller and Frankignoul 1981; Tréguier and Hua 1987). This forcing, and thus the energy input, occurs predominantly at large scales. The energy levels at those scales are therefore higher than in the case of a forcing by the instability of a mean current. This is qualitatively observed on Geosat wavenumber spectra in the northeast Atlantic, which remain red at longer wavelengths (after the spectral peak in wavenumber). Different analyses based on Geosat data led Zlotnicki et al. (1989) to the same conclusions.

The interpretation of wavenumber spectra given by LRB, namely, change in dynamic regime after the spectral wavenumber peak, was confirmed by Le Traon (1991), who performed a global description of time scales and their relationships with space scales in the North Atlantic. Indeed, pseudodispersion relations deduced from Geosat frequency–wavenumber spectra have revealed two distinct dynamic regimes, as in models (e.g., Treguier and Hua 1987): a turbulent regime for smaller scales, where there is proportionality between space and time scales; and a linear regime after the spectral peak in wavenumber where an inverse dispersion relation is generally found. The linear regime is clearly observed in low-energy areas. The corresponding oceanic signal could well be a combination of baroclinic and barotropic Rossby waves directly or indirectly generated by wind fluctuations.

Geosat results thus point to certain characteristics that agree, at least qualitatively, with turbulence models and theories. On the other hand, at high wavenumbers, altimetric spectra fall according to $k^{-\alpha}$, where α is on the order of 2 to 4. This implies isotropic kinetic spectra falling according to $k^{-\alpha+2}$, that is, with slopes of between 0 and -2. This contradicts quasigeostrophic turbulence models that imply slopes of at least -3 (e.g., Hua and Haidvogel 1986; Tréguier and Hua 1987). A possible explanation is that surface variability consists not only of geostrophic mesoscale motions but also smaller-scale ageostrophic motions due to the mesoscale variability of the mixed layer (e.g., Klein and Hua 1988). This small-scale surface variability was clearly observed during the FASINEX experiment (Weller 1991). Depending on its relative influence, it will increase energy seen at shorter wavelengths and make altimeter spectra whiter.

4. Spatial scales

In their characterization of spatial scales, SB note that spatial scales of mesoscale variability are "proportional" to the first internal Rossby radius (RI). Specifically, they find that L0 (first zero crossing of SLA autocorrelation function) and RI follow a linear relationship $L0 = 79.2 + 2.2\ RI$ for the entire Atlantic between 10° and 60° (north and south). However, Le Traon et al. (1991) have shown that the estimation of L0 from altimetry is very sensitive to the polynomial adjustment used to remove orbit error. Similarly, SB note that L1 (integral scale of SLA autocorrelation function) and RI show a relationship $L1 = 40.6 + 1.15\ RI$ for regions polewards of 30° latitude in both hemispheres.

This linear relationship does not mean that mesoscale variability spatial scales and RI are proportional. LRB have thus shown that spatial scales vary by a factor of 2 in the North Atlantic between 20° and 60°N, while internal Rossby radii vary by a factor of 5. Based on results from the MODE and Tourbillon experiments, Mercier and Colin de Verdière (1985), whom SB refer to, have shown that eddy spatial scales and RI in the MODE experiment are both about twice as large as in the Tourbillon experiment. Such a proportionality is clearly not what is observed with Geosat data. This, however, should not be too surprising since altimetric spatial scales take account of the ocean variability for larger wavelengths (>1000 km) and longer periods (>2 yr for Geosat) than for in situ measurements.

The linear relationship between mesoscale variability spatial scales and RI found by SB is interesting. LRB had already noted that altimetric wavenumber spectral peaks occur generally at scales of roughly twice RI. It shows that RI is important to eddy spatial scales, as expected from the dynamics (e.g., Pedlosky 1979) (this does not imply, however, that baroclinic instability is the major mechanism for eddy generation everywhere!). The dynamical significance of the constant in the linear relationship remains, however, obscure.

5. Conclusions

A final comment deals with the comparison of data and models. It is indeed important to compare in situ data with model data, for which altimeter data are very useful. In particular, it is a way of improving models and can also help when interpreting data. SB have used the model results to confirm some of their interpretation and to question the quality of Geosat data in the subtropical areas. Actually, as the energy in the model is less than a tenth of the total observed energy in those areas, using the model as a reference is questionable. More generally, model validation is one of the main reasons for collecting data, and one should be extremely reluctant to reject data that do not agree with model results.

FIG. 1. Mean wavenumber spectra of sea level anomaly deduced from two years of Geosat data between 30° and 40°N in the North Atlantic. Arrows show the spectral peak in wavenumber. Dotted lines correspond to an increase in energy between 50 and 150–200 km (Le Traon et al. 1990).

REFERENCES

Bisagni, J. J., 1989: Wet tropospheric range corrections for satellite altimeter-derived dynamic topographies in the western North Atlantic. *J. Geophys. Res.*, **94,** 3247–3254.

Charney, J. G., 1971: Geostrophic turbulence. *J. Atmos. Sci.,* **28,** 1087–1095.

De Mey, P., 1993: Synoptic estimates of an eddy field in the North Atlantic Current. *Oceanol. Acta*, 537–544.

Frankignoul, C., and P. Müller, 1979: Quasi-geostrophic response of an infinite β-plane ocean to stochastic forcing by the atmosphere. *J. Phys. Oceanogr.*, **9,** 104–127.

Fu, L. L., 1983: On the wave number spectrum of oceanic mesoscale variability observed by the Seasat altimeter. *J. Geophys. Res.*, **88,** 4331–4341.

——, and V. Zlotnicki, 1989: Observing oceanic mesoscale eddies from Geosat altimetry: Preliminary results. *Geophys. Res. Lett.*, **16,** 457–460.

Hua, B. L., and D. B. Haidvogel, 1986: Numerical simulations of the vertical structure of quasi-geostrophic turbulence. *J. Atmos. Sci.*, **43,** 2923–2936.

Jourdan, D., 1990: Observation des structures océaniques par altimétrie satellitaire: Influence des corrections d'environnement sur la restitution du signal mésoéchelle. Ph.D. thesis, Université Paul Sabatier, 171 pp.

——, C. Boissier, A. Braun, and J. F. Minster, 1990: Influence of wet tropospheric correction on mesoscale dynamic topography as derived from satellite altimetry. *J. Geophys. Res.*, **95,** 17 993–18 004.

Klein, P., and B. L. Hua, 1988: Mesoscale heterogeneity of the wind driven mixed layer: Influence of a quasi-geostrophic flow. *J. Mar. Res.*, **46,** 495–525.

Le Traon, P. Y., 1991: Time scales of mesoscale variability and their relationship with spatial scales in the North Atlantic. *J. Mar. Res.*, **49,** 467–492.

——, M. C. Rouquet, and C. Boissier, 1990: Spatial scales of mesoscale variability in the North Atlantic as deduced from Geosat data. *J. Geophys. Res.*, **95,** 20 267–20 285.

——, C. Boissier, and P. Gaspar, 1991: Analysis of errors due to polynomial adjustments of altimeter profiles. *J. Atmos. Oceanic Technol.*, **8,** 385–396.

Mercier, H., and A. Colin de Verdière, 1985: Space and time scales of mesoscale motions in the eastern North Atlantic. *J. Phys. Oceanogr.*, **15,** 171–183.

Monaldo, F., 1990: Path length variations caused by atmospheric water vapor and their effects on the measurement of mesoscale ocean circulation features by a radar altimeter. *J. Geophys. Res.*, **95,** 2923–2932.

Müller, P., and C. Frankignoul, 1981: Direct atmospheric forcing of geostrophic eddies. *J. Phys. Oceanogr.*, **11,** 287–308.

Pedlosky, J., 1979: *Geophysical Fluid Dynamics.* Springer-Verlag, 624 pp.

Rhines, P. B., 1977: The dynamics of unsteady currents. *Marine Modelling. The Sea,* E. D. Goldberg, I. N. McCane, J. J. O'Brien, and J. H. Steele, Eds., Wiley, Vol. 6, 189–318, 1977.

Stammer, D., and C. W. Böning, 1992: Mesoscale variability in the Atlantic Ocean from Geosat altimetry and WOCE high resolution numerical modelling. *J. Phys. Oceanogr.*, **22,** 732–752.

——, H. H. Hinrichsen, and R. H. Käse, 1991: Can meddies be detected by satellite altimetry? *J. Geophys. Res.*, **96,** 7005–7014.

Tréguier, A. M., and B. L. Hua, 1987: Oceanic quasi-geostrophic turbulence forced by stochastic wind fluctuations. *J. Phys. Oceanogr.*, **17,** 397–411.

Weller, R. A., 1991: Overview of the Frontal Air–Sea Interaction Experiment (FASINEX): A study of air–sea interaction in a region of strong oceanic gradients. *J. Geophys. Res.*, **96,** 8501–8516.

Zlotnicki, V., L. L. Fu, and W. Patzert, 1989: Seasonal variability in global sea level observed with Geosat altimetry. *J. Geophys. Res.*, **94,** 17 959–17 969.