content of surface sediments is fairly uniform between 1750 and 2750 m, with accelerated dissolution beginning below that depth. For the mooring experiment, there was a slight increase in dissolution between 2256 and 2869 m, with the level of rapid increase in dissolution in the water column occurring below that depth. These results indicate that within the Panama Basin neither the sedimentary lysocline nor the hydrographic lysocline can be directly related to a transition from saturation to undersaturation. Earlier mooring studies in both the Atlantic (*4, 5*) and the Pacific (*6–8*) also demonstrated a lack of correlation between the hydrographic lysocline and this critical level of carbonate saturation. If the pressure coefficient of $-42.5$ cm$^3$/mole is correct, the results of this study indicate that the dissolution kinetics of biogenic calcite in the laboratory and the deep sea may be similar.

ROBERT C. THUNELL
*Department of Geology* and
*Belle W. Baruch Institute for Marine Biology and Coastal Research, University of South Carolina, Columbia 29208*

ROBIN S. KEIR
SUSUMU HONJO
*Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543*

### References and Notes

1. J. Murray and A. F. Renard, in *Report on the Scientific Results of the Voyage of H.M.S. Challenger During the Years 1872 to 1876* (Longman, London, 1981), p. 1.
2. W. S. Broecker and T. Takahashi, in *The Fate of Fossil Fuel $CO_2$ in the Oceans*, N. R. Andersen and A. Malahoff, Eds. (Plenum, New York, 1977), pp. 213–241.
3. In his Atlantic experiments, Milliman (*4*) used magnesian calcite ooids, planktonic foraminifera, and aragonitic ooids. Honjo and Erez (*5*) used reagent calcite, calcite crystals, synthetic aragonite, planktonic foraminifera, coccoliths, and pteropods. In his Pacific experiment Peterson (*6*) used only spheres of optical calcite, whereas Berger (*7*) used planktonic foraminifera at the same location. In the most recent work of Milliman *et al.* (*8*) in the Pacific, they used planktonic foraminifera, pteropods, and aragonitic ooids.
4. J. D. Milliman, in *The Fate of Fossil Fuel $CO_2$ in the Oceans*, N. R. Andersen and A. Malahoff, Eds. (Plenum, New York, 1977), pp. 641–653.
5. S. Honjo and J. Erez, *Earth Planet. Sci. Lett.* **40**, 287 (1978).
6. M. N. A. Peterson, *Science* **154**, 1542 (1966).
7. W. H. Berger, *ibid.* **156**, 383 (1967).
8. J. D. Milliman, J. Erez, S. Honjo, in preparation.
9. W. H. Berger, in *Dissolution of Deep-Sea Carbonates*, W. V. Sliter, A. W. H. Bé and W. H. Berger, Eds. (Special Publication 13, Cushman Foundation for Foraminiferal Research, Washington, D.C., 1975), pp. 82–86. The hydrographic lysocline is a level of accelerated dissolution in the water column, as originally observed by Peterson (*6*). This is different from both the sedimentary lysocline, which is the level of decreased carbonate content on the sea floor, and the foraminiferal lysocline, which is the level where there is a drastic change in the preservation of planktonic foraminifera.
10. W. H. Berger, *Deep-Sea Res.* **15**, 31 (1968); *Mar. Geol.* **8**, 111 (1970); F. L. Parker and W. H. Berger, *Deep-Sea Res.* **18**, 73 (1971); V. Kolla, A. W. H. Bé, P. E. Biscaye, *J. Geophys. Res.* **81**, 2605 (1976); W. H. Berger, C. G. Adelseck, L. A. Mayer, *ibid.*, p. 2617; P. E. Biscaye, V. Kolla, K. K. Turekian, *ibid.*, p. 2595.
11. J. W. Morse and R. A. Berner, *Am. J. Sci.* **272**, 840 (1972).
12. W. S. Broecker and J. Takahashi, *Deep-Sea Res.* **25**, 65 (1978).
13. J. M. Edmond, *ibid.* **21**, 455 (1974).
14. J. Erez and S. Honjo, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, in press.
15. $\Omega = [Ca^{2+}]\,[CO_3^{2-}]/K_{sp}'$, where $[CO_3^{2-}]$ and $[Ca^{2+}]$ are the in situ dissolved total $CO_3^{2-}$ and $Ca^{2+}$ concentrations, and $K_{sp}'$ is the in situ apparent solubility product for calcite. This index is calculated in the following way (i) Total $CO_2$ is calculated from the *p*H and alkalinity measurement at 1 atm and 35°C and the in situ $[CO_3^{2-}]$ is determined from the total $CO_2$ and titration alkalinity [R. S. Keir, *Mar. Chem.* **8**, 95 (1979)]. (ii) The value of $K_{sp}'$ at 1 atm is calculated from the *p*H and alkalinity measurements of S. E. Ingle, C. Culberson, J. Hawley, and R. Pytkowicz [*Mar. Chem.* **1**, 295 (1973)]. (iii) Lyman's [thesis, University of California, Los Angeles (1956)] $K_1'$, $K_2'$, and $K_B'$ ($K_1'$ and $K_2'$ are the first and second dissociation constants for carbonic acid and $K_B'$ is the first dissociation constant for boric acid) are used to calculate total $CO_2$ and $K_{sp}'$ at 1 atm, and the pressure corrections of C. H. Culberson and R. Pytkowicz [*Limnol. Oceanogr.* **13**, 403 (1968)] are applied to $K_1'$, $K_2'$, and $K_B'$ in order to calculate in situ $[CO_3^{2-}]$. Two pressure coefficients, $-36.0$ cm$^3$/mole (*12, 20*) and $-42.5$ cm$^3$/mole (*21*), at 0°C are used to correct the 1-atm $K_{sp}'$ to in situ pressure.
16. T. C. Moore, Jr., G. R. Heath, R. O. Kowsman, *J. Geol.* **81**, 458 (1973).
17. G. M. Friedman, *Geol. Soc. Am. Bull.* **76**, 1191 (1965).
18. C. H. Culberson, thesis, Oregon State University (1972).
19. B. Wyatt, W. Gilbert, W. Gordin, D. Barstow, *Data Report No. 42* (Reference 70-12, Office of Naval Research, Oregon State University, Corvallis, 1970).
20. R. H. Pytkowicz and G. A. Fowler, *Geochem. J.* **1**, 169 (1967).
21. S. E. Ingle, *Mar. Chem.* **3**, 301 (1975); T. Takahashi, in *Dissolution of Deep-Sea Carbonates*, W. V. Sliter, A. W. H. Bé, W. H. Berger, Eds. (Special Publication 13, Cushman Foundation for Foraminiferal Research, Washington, D.C., 1975), pp. 11–26.
22. J. K. B. Bishop, R. Collier, D. R. Ketten, J. Edmond, *Deep-Sea Res.*, in press.
23. S. A. Swift and C. Wenkam, *Mar. Geol.* **27**, 67 (1978).
24. R. A. Berner and P. Wilde, *Am. J. Sci.* **272**, 826 (1972).
25. R. A. Berner, *ibid.* **276**, 713 (1976).
26. R. S. Keir, *Geochim. Cosmochim. Acta* **44**, 241 (1980). The laboratory dissolution rate curve is based on measurements made on planktonic foraminifera (larger than 300 μm) collected from a water depth of 1500 m on the Ontong-Java Plateau.
27. We thank J. Erez for discussions and suggestions. P. G. Brewer, chief scientist of R.V. *Knorr* leg 73, kindly let R.S.K. acquire carbonate hydrographic data during his Panama Basin cruise. This study was supported in part by NSF contracts OCE77-27080 and OCE80-23959. Contribution 371 from the Belle W. Baruch Institute for Marine Biology and Coastal Research. Contribution 4638 from the Woods Hole Oceanographic Institution.

9 June 1980; revised 18 November 1980

# Method for Estimation of Ocean Current Velocity from Satellite Images

Abstract. *Barotropic instability waves on a shear interface propagate at the average speed of the water on the two sides. Assuming the instability to be excited by tidal oscillations, the phase speed is the wavelength divided by the tidal period. If the water is at rest on one side of the shear layer the current speed on the other side can be calculated. This method, applied to the Gulf Stream beyond Cape Hatteras as seen in satellite images, gives estimates of current speed in general agreement with in situ observations.*

Infrared satellite images show the sharp thermal boundaries formed where ocean currents separate from topographic boundaries. These thermal interfaces are also velocity interfaces (free shear layers), which in turn are unstable to perturbations caused by disturbances originating elsewhere in the ocean. A clear example of such a free shear layer is that formed by the inshore edge of the Gulf Stream as it passes Cape Hatteras, North Carolina, as shown in Fig. 1. The image is a Tiros-N infrared image with resolution ~ 1 km, obtained on digital tape from the Environmental Data and Information Service of the National Oceanic and Atmospheric Administration and processed further in our facilities at Massachusetts Institute of Technology for navigational reference and contrast enhancement.

The undulations in the free shear layer beyond Cape Hatteras contain a strongly periodic component. The apparently periodic waves upstream of Cape Hatteras we attribute to a local topographic disturbance. Figure 1 shows the inshore and offshore maxima in shear layer excursion marked with short line segments. The nearly periodic structure of the shear instability suggests that its excitation is periodic. Laboratory experiments by Browand (*1*) and Miksad (*2*) showed that periodic excitation of a free shear layer organized the instability fluctuations in a nearly periodic structure, while random excitation gave a random structure. Figure 2 shows the power spectral densities of velocity fluctuations near a free shear layer that is unstable; the spectra are shown for a series of distances from the origin of the shear layer (*2*). The upper part of Fig. 2 shows spectra with random, nearly white-noise excitation, while the lower part shows the result for periodic excitation.

  

One is led to conclude that periodic fluctuations in a free shear layer are most likely caused by periodic excitation. We suggest that the periodic shear instability structure in the edge of the Gulf Stream, shown in Fig. 1, is due to periodic excitation, and that the excitation is the fluctuations associated with the semidiurnal lunar ($M_2$) tide, which has a period of approximately 12.4 hours.

When one knows both the wavelength and the period of a propagating disturbance, one finds the propagation speed as the ratio, and we find the average speed of the first few shear waves to be 1.2 m/sec (2.4 knots).

A further assumption is that the instability is barotropic, so that for a layer that is thin compared to the instability wavelength, the propagation speed of the instability is the average of the speed of the water on the two sides. By the speed of the water we mean the average speed over many fluctuations. Stommel (*3*) showed that the Gulf Stream shear layer is thin beyond Cape Hatteras. If we now take the average speed of the water inshore of the shear layer to be negligible, we can calculate the average speed of the water on the offshore side of the shear layer, and we find that the average current there is 2.4 m/sec (4.8 knots).

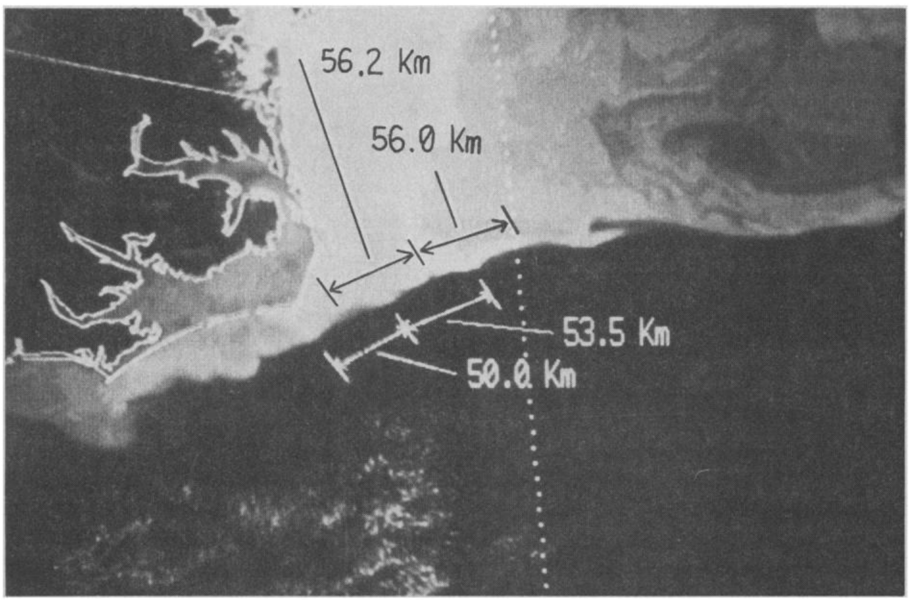


Fig. 1. Enhanced infrared image of the sea surface near Cape Hatteras, North Carolina, showing shear instability waves, with wavelengths indicated, in the Gulf Stream edge. The image was constructed from Tiros-N data taken on 23 March 1979. The dotted, nearly vertical line is the satellite nadir track; the coast and the state boundary are also superimposed on the image.

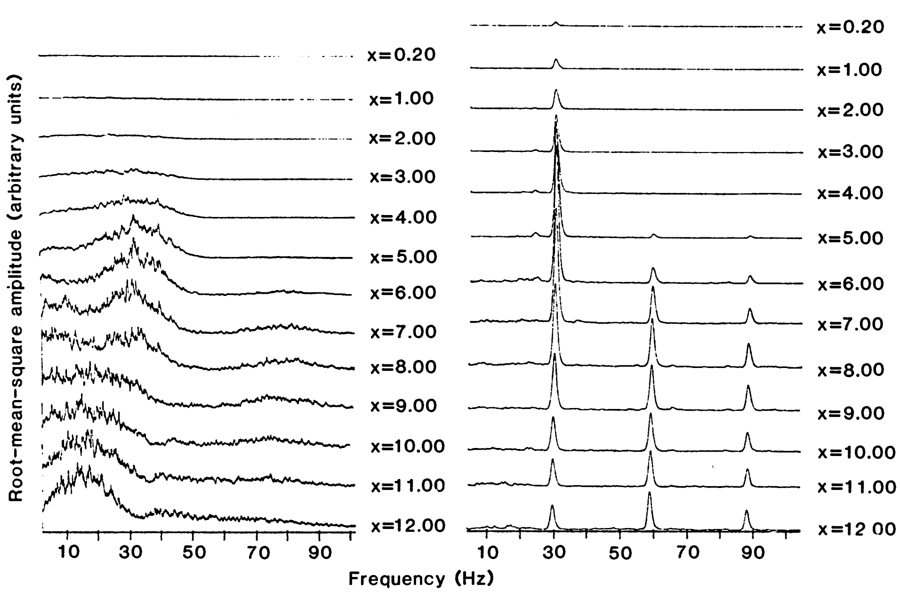


Fig. 2. Velocity fluctuations measured in a free shear layer in the laboratory, at a series of positions downstream from the origin. The upper half represents random excitation, the lower half periodic excitation. Note the line spectrum that indicates a periodic response.

Boisvert (*4*) found a current beyond the offshore edge of the shear layer to be close to 4 knots, as a maximum value, while the minimum observed current was much less [see also Richardson *et al.* (*5*)]. Most current observations are from fixed current meters. A current meter at the mean position of the Gulf Stream edge would be inside the current part of the time and outside part of the time. The observed average current would then be less than the actual current.

Our test of the method of estimating current from a satellite image, by assuming tidal excitation and barotropic instability, thus yields results in general agreement with observations. The method can, of course, be refined by use of in situ observations and of observational results for other shear layers farther inshore, if they can be found. Preliminary investigation indicates that the method gives good results when used to estimate the flow south of Georges Bank, in the Florida Current at Grand Bahama Island, as well as in other places. Although the method requires further verification and refinement, we feel that it may become a useful tool for current speed estimates in places that are inaccessible or inconvenient for ships and buoys, and that it may be able to provide long-term time series of a crude kind when in situ observations contain gaps in location and time.

ERIK MOLLO-CHRISTENSEN
PETER CORNILLON
AFFONSO DA S. MASCARENHAS, JR.
*Department of Meteorology and Physical Oceanography, Massachusetts Institute of Technology, Cambridge 02139*

**References and Notes**

1. F. K. Browand, *J. Fluid Mech.* **26**, 281 (1966).
2. R. W. Miksad, *ibid.* **56**, 695 (1972).
3. H. Stommel, *The Gulf Stream* (Univ. of California Press, Berkeley, 1958).
4. W. E. Boisvert, "Major currents in the North and South Atlantic oceans between 64°N and 60°S" (Technical Report TR-193, Naval Oceanographic Office, Washington, D.C., 1967).
5. W. S. Richardson, W. J. Schmitz, Jr., P. P. Niller, *Deep-Sea Res.* **16**, 225 (1969).
6. Supported by the Office of Naval Research under contracts N00014-79-C-0413 and N00014-75-C-0291. A.S.M. was supported by a fellowship from the National Research Council of Brazil. The image processing software was developed by R. Evans, O. Brown, and J. Brown of the University of Miami with Office of Naval Research funds. The help of R. Evans and J. Brown in installing the software package at Massachusetts Institute of Technology is gratefully acknowledged. K. A. Morey and R. Thayer were instrumental in setting up the processing system. P.C. is on leave from the University of Rhode Island; A.S.M. is on leave from the Oceanographic Institute, University of São Paulo, Brazil.

29 October 1980