The Shaping of Continental Slopes by Internal Tides

D. A. Cacchione,[1] L. F. Pratson,[2] A. S. Ogston[3]

The angles of energy propagation of semidiurnal internal tides may determine the average gradient of continental slopes in ocean basins (~2 to 4 degrees). Intensification of near-bottom water velocities and bottom shear stresses caused by reflection of semi-diurnal internal tides affects sedimentation patterns and bottom gradients, as indicated by recent studies of continental slopes off northern California and New Jersey. Estimates of bottom shear velocities caused by semi-diurnal internal tides are high enough to inhibit deposition of fine-grained sediment onto the slopes.

Continental slopes are the steepest parts of continental margins, starting at the edge of continental shelves in water depths of about 100 to 200 m and extending to the upper continental rise in water depths of about 1500 to 3000 m. The average slope of continental slopes is about 2° to 4°, an order of magnitude lower than the internal angle of repose of the marine muds that make up slopes (*1, 2*).

Internal waves and tides are energetic, ubiquitous, and persistent oceanic phenomena. When internal waves reflect from a sloping bottom, the frequencies of the incident waves are preserved upon reflection; reflection of energy rays or characteristics occurs about the direction of the local gravity vector (not about the direction normal to the slope) (*3*). Characteristics of internal tidal energy can approach the slope γ from the ocean interior (Fig. 1) or from local generation areas at the shelf-slope boundary (*4*). Three reflection conditions for internal waves normally incident onto a slope-shelf are possible (Fig. 1, B to D).

The internal tidal characteristics (Fig. 1, A to D) travel at an angle c to the horizontal that is determined by the frequency of the internal waves, the density profile, and latitude (*3*). This angle is given by

$$c = \left(\frac{\sigma^2 - f^2}{N^2 - \sigma^2}\right)^{1/2} \qquad (1)$$

where σ is the internal wave frequency [σ = 0.081 cycles per hour (cph) for internal tides at semi-diurnal tidal frequency], and ϕ is the local inertial frequency at latitude ϕ (in degrees); $f = (\sin\phi)/12$ cph, and its value ranges from zero at the equator to 0.083 cph at the poles. N is the Brunt-Vaisala frequency (or stability frequency) and is usually calculated from the vertical oceanic density gradient $\partial\rho/\partial z$ according to $N = [(g/\rho)(\partial\rho/\partial z)]^{1/2}$ s^{-1} or $N = \sim 573[(g/\rho)(\partial\rho/\partial z)]^{1/2}$ cph, where ρ is density and g is gravitational acceleration (~9.8 m/s^2). Typical values of N in the ocean at continental slope depths from about 200 to 2000 m are $1 < N < 4$ cph (*3*). Much of the vertical density structure in the oceans below 200 m can be approximated by a linear gradient, as shown in Fig. 1A.

Calculations of γ/c were made for the New Jersey and northern California continental slopes (*5, 6*). Bottom gradients γ were computed from high-resolution bathymetric data (*7*) and averaged over 500-m^2 grid areas. The estimates of c were based on seasonally averaged vertical profiles of N that had been derived from historical density data (*8*). Characteristic angles c for semi-diurnal internal tides were calculated from Eq. 1 at fixed depth intervals and interpolated at each grid point to compute γ/c.

Bottom velocities and bottom shear stresses should be highest where the slopes are critical ($\gamma/c = 1$). On the New Jersey continental slope (Fig. 2B), large regions between 200 and 2000 m are critical during winter (Fig. 2D). This includes the interior sections of many of the submarine canyons, such as the central portion of Hudson Submarine Canyon (large canyon in the upper right of Fig. 2D) where strong internal tidal velocities have been measured (*9*).

During summer, much of the continental slope is reflective ($\gamma/c > 1$) and bottom velocities should be small (Fig. 2C, bottom). However, the upper slope and upper parts of canyon heads are critical or near-critical, and the entire section of the slope southeast of Hudson Submarine Canyon (also called the Hudson Apron) is critical. This relatively large area may have high internal tidal bottom velocities and could be a region where internal tides are generated (*4*). The strong seasonal contrasts in γ/c are caused by seasonal variations in the vertical density profiles.

Unlike the New Jersey continental slope, the northern California slope shows less seasonal variation in the geographic distribution of γ/c (Fig. 2C). Seasonal contrasts in the vertical density profiles (and N) are smaller in this region than off New Jersey below about 200 m water depths (*8*). Over much of this part of the upper continental slope, $\gamma/c < 1$. Around the mooring at 450 m depth (Fig.

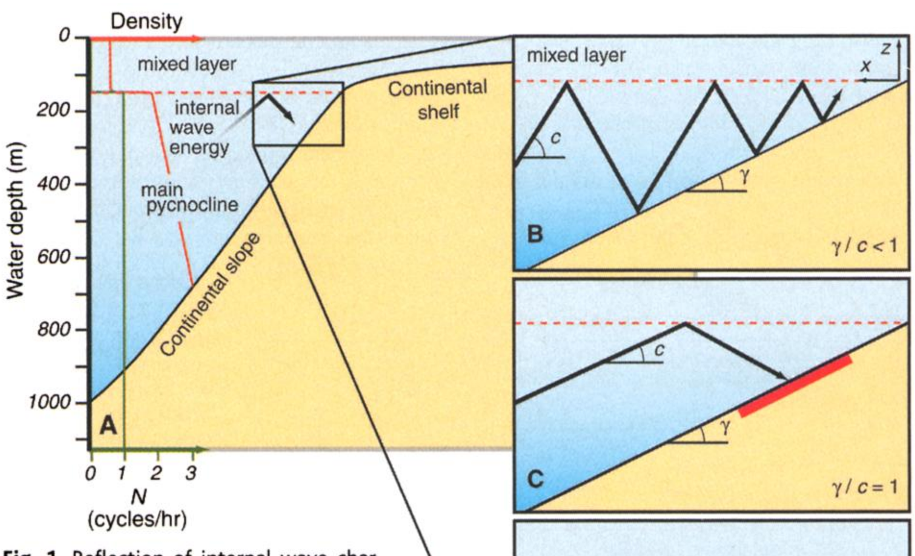

Fig. 1. Reflection of internal wave characteristics (black lines) from bottom slope γ for (**A**) simplified density profile (red) and Brunt-Vaisala frequency profile N (green), transmissive [(**B**), $\gamma/c < 1$], critical [(**C**), $\gamma/c = 1$], and reflective [(**D**), $\gamma/c > 1$]. Energy is trapped along the bottom, and bottom velocities are intensified in (C) (maroon bar indicates velocity intensification); bottom velocities also increase upslope in (B) (*10*).

[1]Coastal and Marine Geology MS-999, U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025, USA. [2]Earth & Ocean Sciences, Nicholas School of the Environment and Earth Sciences, Duke University, Durham, NC 27708, USA. [3]University of Washington, Box 357940, Seattle, WA 98195, USA.

2C), the slope is near critical ($\gamma/c \sim 0.8$).

Long-term measurements (1995 to 1999) of currents, temperatures, salinities, and suspended sediment concentrations were made on the northern California slope in about 450 m water depth (Fig. 2C) (*5*). These data were obtained along a mooring at 15, 270, and 390 m above the bottom. Ten additional temperature sensors were mounted on the mooring to gather hourly temperature data between 15 and 300 m above the bottom. A 3-day section of the mooring data shows that the highest near-bottom flow speeds reached about 40 cm/s on 11 April 1997 and were directed downslope (Fig. 3A). The downslope tidal currents were typically of longer duration than upslope currents, producing net downslope low-frequency flows (Fig. 3A).

Over the same 3-day period, the temperature records at about 15, 50, and 100 m above the bottom showed coalescing isotherms that may have been caused by strong near-bottom advection and mixing (Fig. 3B). The isotherm variations suggest that a thermal bore of tidal frequency with sharp horizontal temperature gradients moved the colder water upslope, followed by a more prolonged downslope movement of warmer water. The steep temperature drop during the latter half of 12 April 1997 (Fig. 3B) was associated with strongly ac-

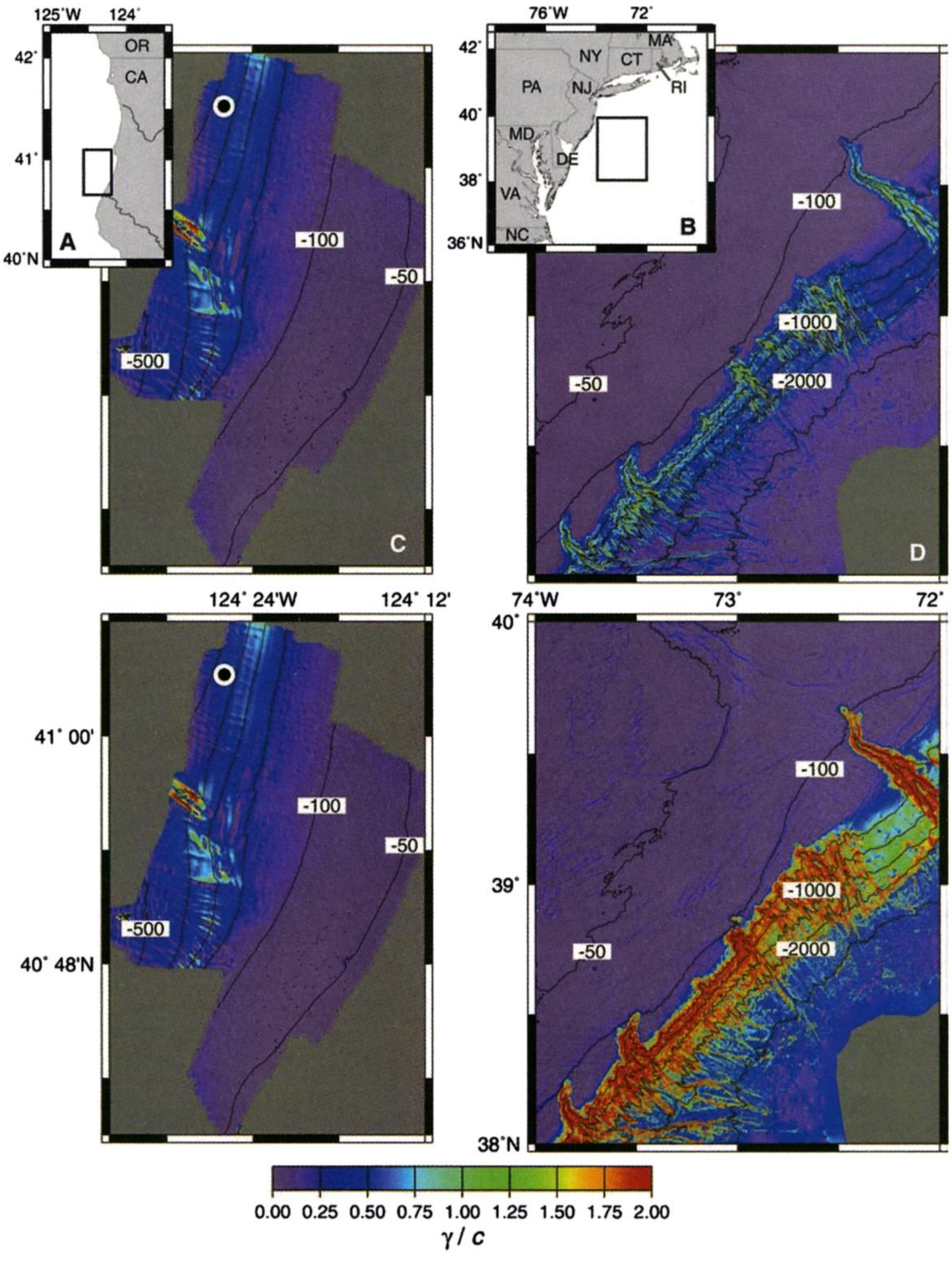

Fig. 2. (**A** and **B**) Two contrasting shelf and slope study areas were evaluated for internal wave reflection conditions, off northern California (A) and off New Jersey (B). (**C** and **D**) Critical slope conditions for semi-diurnal internal tide are shown in green for mean winter (top) and summer (bottom) vertical density profiles. Water-depth contours are in meters. A current meter mooring was anchored in about 450 m of water off northern California (*5*) (white and black circles).

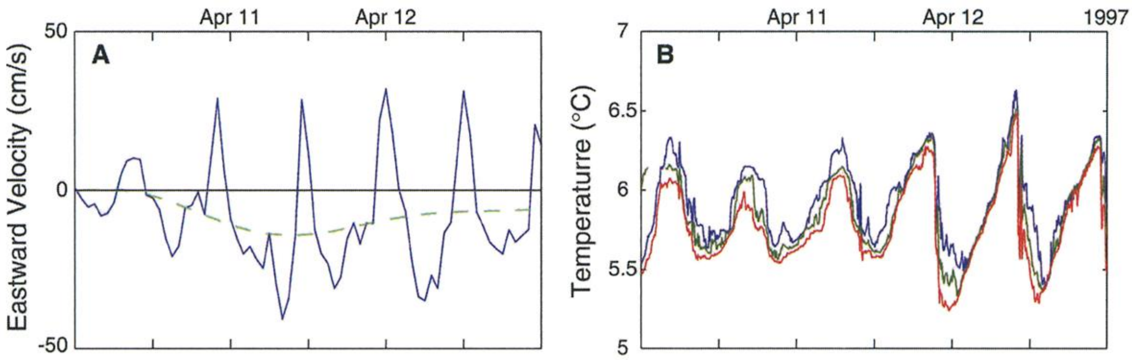

Fig. 3. Across-slope currents measured at 15 m above the bottom (**A**); temperatures measured at 15 (red), 50 (green), and 100 (blue) m above bottom (**B**). Dashed line in (A) indicates low-frequency current.

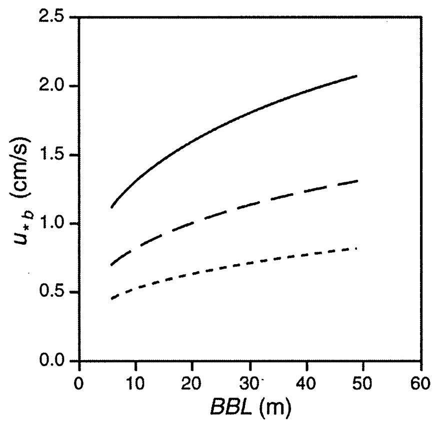

Fig. 4. Bottom velocity shear u_{*b} plotted against thickness of bottom boundary layer (BBL) h using Eq. 2. Solid, long-dashed, and short-dashed lines indicate u_o = 20, 10, and 5 cm/s, respectively.

celerative upslope flow during which the velocity changed from about 30 cm/s downslope to about 35 cm/s upslope over a period of about 6 hours (Fig. 3A). Turbulent mixing caused by internal tidal bores could inhibit sedimentation on continental slopes (*10, 11*) or, in the case of obliquely incident internal tidal bores, could transport sediment along the slope to other lateral depocenters (*12*).

Deposition of fine-grained sediment onto the seabed can occur only if the bed shear stress τ_b is less than the limiting "depositional" bed stress τ_d. Field and laboratory results (*13*) suggest that for fine-grained sediment $0.6 < \tau_d < 1.0$ dynes/cm^2 (or equivalently, $0.7 < u_{*d} < 0.9$ cm/s), where τ is expressed in terms of a shear velocity u_* ($\tau = \rho u_*^2$).

Similarly, erosion of fine-grained cohesive sediment occurs when $\tau_b > \tau_e$, a critical or threshold stress required to erode the bottom sediment. Typical values for loosely consolidated cohesive sediment are $1.0 < \tau_e < 2.0$ dynes/cm^2 ($0.9 < u_{*e} < 1.4$ cm/s) (*14*). Therefore, approximate threshold conditions for deposition and erosion of fine-grained cohesive sediment on continental slopes are about $u_{*d} < 1.0$ cm/s and $u_{*e} > 1.0$ cm/s, respectively.

Threshold conditions for deposition and erosion of slope sediment can be compared to expected estimates of bottom friction velocity u_{*b} for critical and near-critical internal tides. For critical internal waves over a slope, u_{*b} can be obtained by assuming the equivalence of dissipation and production of turbulent kinetic energy in the bottom boundary layer. This assumption is only approximate and is made here for cycle-averaged quantities (no time dependence is considered in this simplified model). Equating the cycle-averaged bottom dissipation rate for critical and near-critical internal waves (*15*) and the average rate of production of energy per unit mass of fluid (*16*) yields

$$u_{*b} = M u_o^{2/3} \qquad (2)$$

$$M = \left[\frac{0.16h(1-r)}{\ln(h/z_o)} N \cos\alpha \sin 2(\alpha+\beta)\right]^{1/3} \qquad (3)$$

where the angles α and β represent characteristic and bottom slope angles, respectively ($\alpha = \tan^{-1} c$ and $\beta = \tan^{-1}\gamma$), h is thickness of the internal tide bottom boundary layer, z_o is the hydraulic roughness parameter, r is the reflection coefficient, and u_o is internal wave input velocity.

With a typical oceanic value for $N = 1.0$ cph and an average slope angle of $\alpha = 2.5°$, Eqs. 2 and 3 indicate that u_{*b} increases with both u_o and h (Fig. 4). Although no definitive field studies have determined h for critical internal tides over continental slopes, conductivity-temperature-depth (CTD) profiles spanning a variety of locations and water depths suggest that bottom mixed layers of 10 to 50 m are common on continental slopes (*17*). Off northern California, the bottom mixed layers based on CTD profiles were typically 20 to 40 m thick (*18*). Assuming h = 20 m, u_o = 10 cm/s, z_o = 1 cm, and r = 0.3 (a conservative value), $u_{*b} \sim 1$ cm/s (Fig. 4). This result suggests that for $h > 20$ m and $u_o > 10$ cm/s, u_{*b} exceeds u_{*d}, deposition of fine-grained sediment would be inhibited, and erosion of bottom sediment might occur. Over many continental slopes, h and u_o are probably greater than the values used in this example, and the bottom shear velocities would be larger.

Our analysis indicates that under typical oceanic conditions, critical semi-diurnal internal tides reflecting on continental slopes can generate bottom shear velocities that are high enough to inhibit deposition of fine-grained sediment and possibly erode the sediment surface. Near-critical conditions like those on the northern California upper continental slope produced high (40 cm/s) near-bottom internal tidal velocities and internal tidal bores that might keep fine-grained sediment from depositing. These episodically high near- bottom velocities (and associated high bottom shear stresses) probably account for the relatively low sediment accumulation rates and slightly larger surficial bottom sediment sizes that were reported for this section of the slope (*19*).

What has not yet been considered is the impact of this process on continental slopes over geologic time. Our findings suggest that at least sections of the world's continental slopes appear to be in equilibrium with the internal tide energy. How this equilibrium was established, and over what time periods, is as yet challenging to conceptualize. One possibility is that shelf-slope strata steepened until reaching the critical angle while prograding across the continental-oceanic crust transition. Before this point, bottom shear stresses caused by the internal tides would be low enough to allow the deposition on the upper slope, driving the slope steepening. However, as the slope approached the critical angle, these bottom shear stresses would have become high enough to inhibit deposition and possibly even erode the upper slope. This change could arrest further slope steepening, and over geologic time might eventually lead to a regional slope gradient that is in equilibrium with the semi-diurnal internal wave field.

Of course, this argument presupposes that the sediments that make up continental slopes were initially laid down at angles equal to or less than the critical angle. If they were laid down on greater slopes, the deposits could eventually become even steeper because the internal tidal energy decreases with increasing seafloor gradient above the critical angle. Other processes are at work in shaping continental slopes, and at least one of these, turbidity currents, also has the potential to be an important cause for the relative flatness of continental slopes (*20*).

References and Notes

1. L. F. Pratson, W. F. Haxby, *Geology* **24**, 3 (1996).
2. D. O'Grady, J. Syvitski, L. F. Pratson, J. F. Sarg, *Geology* **28**, 207 (2000).
3. O. M. Phillips, *The Dynamics of the Upper Ocean* (Cambridge Univ. Press, London, 1977).
4. S. J. Prinsenberg, W. Wilmot, M. Rattray Jr., *Deep-Sea Res.* **16**, 179 (1974).
5. D. A. Cacchione, L. Pratson, A. Ogston, paper presented at the Ocean Sciences Meeting of the American Society of Limnology and Oceanography and the American Geophysical Union, San Antonio, TX, 24 to 28 January 2000.
6. Long-term studies of sediment processes and the formation of stratigraphic sequences were carried out on the continental terraces off New Jersey and northern California during the U.S. Office of Naval Research (ONR) STRATAFORM program from 1995 to 1999. The continental slopes in these two regions have different regional gradients (4° in New Jersey versus 2° in northern California) and different geology (e.g., New Jersey is a passive margin and northern California is an active margin).
7. L. Pratson *et al.*, *Eos* **80**, 37 (1999).
8. M. E. Conkright *et al.*, *Nat. Ocean. Data Ctr. Int. Rep. 14* (1999).
9. F. Hotchkiss, C. I. Wunsch, *Deep-Sea Res.* **29**, 415 (1982).
10. D. A. Cacchione, C. I. Wunsch, *J. Fluid Mech.* **66**, 223 (1974).
11. J. Ribbe, P. Holloway, *Cont. Shelf Res.* **21**, 395 (2001).
12. O. Zikanov, D. N. Slinn, *J. Fluid Mech.* **445**, 235 (2001).
13. I. N. McCave, in *Shelf Sediment Transport: Process and Pattern*, D. Swift, D. Duane, O. Pilkey, Eds. (Dowden, Hutchinson, and Ross, Stroudsburg, PA, 1972), chap. 10.
14. R. J. S. Whitehouse, R. L. Soulsby, W. Roberts, H. J. Mitchener, *Dynamics of Estuarine Muds* (Report SR 527, HR Wallingford Ltd., Wallingford, UK, 1999).
15. G. N. Ivey, K. B. Winters, I. P. D. DeSilva, *J. Fluid Mech.* **418**, 59 (2000).
16. J. S. Turner, *Buoyancy Effects in Fluids* (Cambridge Univ. Press, Cambridge, ed. 1, 1973), pp. 128–130.
17. R. R. Dickson, I. N. McCave, *Deep-Sea Res.* **33**, 791 (1986).
18. E. E. McPhee, thesis, University of Washington, Seattle (2000).
19. C. R. Alexander, A. M. Simoneau, *Mar. Geol.* **154**, 243 (1999).
20. G. Parker, S. Kostic, J. Marr, *J. Sed. Res.*, in press.

21. We thank the ONR Marine Geology and Geophysics Program and J. Kravitz for supporting this work in STRATAFORM. Additional support came from ONR Physical Oceanography and U.S. Geological Survey's Coastal and Marine Program.

14 January 2002; accepted 21 March 2002

Global Cooling After the Eruption of Mount Pinatubo: A Test of Climate Feedback by Water Vapor

Brian J. Soden,[1]* Richard T. Wetherald,[1] Georgiy L. Stenchikov,[2] Alan Robock[2]

The sensitivity of Earth's climate to an external radiative forcing depends critically on the response of water vapor. We use the global cooling and drying of the atmosphere that was observed after the eruption of Mount Pinatubo to test model predictions of the climate feedback from water vapor. Here, we first highlight the success of the model in reproducing the observed drying after the volcanic eruption. Then, by comparing model simulations with and without water vapor feedback, we demonstrate the importance of the atmospheric drying in amplifying the temperature change and show that, without the strong positive feedback from water vapor, the model is unable to reproduce the observed cooling. These results provide quantitative evidence of the reliability of water vapor feedback in current climate models, which is crucial to their use for global warming projections.

Water vapor plays a key role in regulating Earth's climate. It is the dominant greenhouse gas (*1*) and provides the largest known feedback mechanism for amplifying climate change (*2*). Because the equilibrium vapor pressure of water increases rapidly with temperature, it is generally believed that the concentration of water vapor will rise as the atmosphere warms. If so, the added radiative absorption from water vapor will act to further amplify the initial warming. Current climate models suggest that this provides an important positive feedback, roughly doubling the sensitivity of the surface temperature to an increase in anthropogenic greenhouse gases (*3–5*). If the actual feedback by water vapor is substantially weaker than predicted by current models, both the magnitude of warming and range of uncertainty resulting from a doubling of CO_2 would be substantially diminished (*5*).

Despite the importance of water vapor feedback in determining the sensitivity of Earth's climate, the fidelity of its representation in climate models has remained a topic of debate for more than a decade (*6*, *7*). The difficulty in verifying models partly stems from the lack of observed climate variations that can provide quantitative tests of the feedbacks in question. Assessments of water vapor feedback are often based on regional, seasonal, or interannual variations of Earth's climate (*8–12*), which differ markedly in both cause and character from the more uniform, radiatively forced perturbations that result from increasing CO_2. Thus, their conclusions are often qualitative, and their relevance to feedbacks that arise from global warming are often questioned (*6*, *13–15*).

It has long been recognized that volcanic eruptions provide a valuable opportunity to observe the climate system's response, albeit a transient one, to the presence of an external radiative forcing (*16–19*). Strong volcanic eruptions inject large amounts of sulfuric gas into the lower stratosphere where it combines with water and oxygen to form small, yet optically important, aerosol particles. Winds rapidly disperse the particles throughout the lower stratosphere, resulting in a near-global perturbation to the radiative energy balance. Because they are more effective at scattering sunlight than absorbing longwave terrestrial radiation, the net radiative effect of volcanic aerosols is to cool the planet.

The eruption of Mount Pinatubo in the Philippines in June of 1991 resulted in unprecedented observations of both radiative forcing from volcanic aerosols as well as the climate system's response to this forcing. Satellite observations confirm the decrease in solar heating due to Mount Pinatubo aerosols (*20–22*), which led to a global cooling of the lower troposphere (*23*, *24*). Associated with this cooling was a reduction in the global water vapor concentrations, which closely tracked the decrease in temperature (*25*). Thus, Mount Pinatubo provides a unique opportunity to not only study the sensitivity of the climate system but, more importantly, to also assess the response of water vapor and quantify its role in determining that sensitivity.

It is widely recognized that current climate models possess a strong positive feedback by water vapor (*3–5*, *26*). Nevertheless, one can

[1]Geophysical Fluid Dynamics Laboratory/National Oceanic and Atmospheric Administration, Princeton University, Princeton, NJ 08542, USA. [2]Department of Environmental Sciences, Rutgers University, New Brunswick, NJ 08901, USA.

*To whom correspondence should be addressed. E-mail: bjs@gfdl.gov