Gravity waves in the lower atmosphere, forced by flow over mountains, have been observed and modeled for many years (1). Mountain waves can also propagate into the stratosphere. Drag produced by breaking stratospheric mountain waves is essential for accurate climate and forecast models (2). The turbulence generated by stratospheric mountain wave breaking poses a major safety hazard to high-altitude aircraft (3–5) and influences stratosphere-troposphere exchange of important trace gases (6). Mountain wave ascent triggers formation of polar stratospheric clouds (PSCs) (7). It has been argued that mountain wave PSCs reduce Arctic ozone concentrations by activating chlorine in stratospheric air that flows through them (8).

Computational limits have prevented global troposphere-stratosphere models from operating at a spatial resolution fine enough to generate mountain waves in the stratosphere, so their effects must be parameterized (2). The spatial resolution of stratospheric sounding satellites has generally been too coarse to resolve mountain waves. Detailed measurements have been limited largely to scattered ground-based soundings (9), some regional campaigns (3, 10), and chance intercepts by high-altitude aircraft (4, 11). Thus, global data are lacking.

Recently, a few high-resolution stratospheric satellite instruments have detected gravity waves (12). However, instrumental effects have complicated interpretation of these gravity-wave signals (13). In particular, the data have shown little clear evidence of mountain waves, despite suggestions that mountains should be a strong source of stratospheric gravity waves (2–4, 9–11).

On 4 November 1994, the CRISTA-SPAS (Shuttle Pallet Satellite) experiment was deployed into orbit by the space shuttle Atlantis (Fig. 1A). CRISTA measured infrared emission spectra (4 to 71 μm) from atmospheric limb scans with the use of three telescopes that viewed closely separated volumes of air (14). From the CO2 Q-branch emission at ~12.6 μm, vertical temperature profiles (z) with a precision of ~0.5 K were derived every 1.5 km in altitude, z, throughout the stratosphere (15, 16). Although short spatial scales along the line of sight are smeared out, medium scale structure is resolved. Theoretical radiance calculations show that the temperature data have >50% visibility to gravity waves with wavelengths ≥3 to 5 km vertically and ≥100 to 200 km along the limb (17). Each telescope provided a profile every 200 or 400 km along the orbital track, depending on the measurement mode (18), for a week.

Small-scale temperature fluctuations z(19) were isolated from global temperature data with a wavenumber 0 to 6 Kalman filter. Vertical wavelengths l and amplitudes z of dominant oscillations in each z profile were determined with spectral peak detection and sliding harmonic fits, respectively. These parameters are shown (Fig. 1B) for three successive z profiles (labeled 1, 2, and 3) acquired over the southern Andes Mountains (Fig. 1C) on 6 November 1994. All three profiles show similar-looking large-amplitude wave oscillations at 15 to 30 km. The photo in Fig. 1A, taken over this region 2 days earlier, shows hatched mountain wave clouds downstream of the Andes. This suggests that the oscillations in Fig. 1B may be stratospheric mountain waves.

A long, stationary plane mountain wave emanating from the Andes should have a vertical wavelength given by the hydrostatic relation

\[ \lambda_{\text{theory}} \approx \frac{2\pi U}{N} \]  

(1)

where \( U \) is the local horizontal wind speed across the Andean Ridge and \( N \) is the buoyancy frequency (1). At 15 to 30 km, \( U \approx 20 \) to 23 m s\(^{-1}\) (Fig. 1D) and \( N \approx 0.020 \) rad s\(^{-1}\), which yields \( \lambda_{\text{theory}} \approx 6 \) to 7 km. The observed vertical wavelength \( \lambda \) of the measured wave oscillations in Fig. 1B is ~6.5 km at 15 to 30 km. The close agreement between observed and theoretical vertical wavelengths is strong evidence that these measured wave oscillations are stratospheric mountain waves. Successive profiles are separated horizontally by 200 km (Fig. 1C) and phase shifted by 180° (Fig. 1B). This implies a horizontal wavelength \( \lambda_h = 400 \) km for this wave, where \( n \) is an odd integer. Given the sensitivity of CRISTA to \( \lambda_h > 100 \) to 200 km, then \( \lambda_h \) must be either 130 or 400 km.

Nonbreaking mountain waves grow in amplitude with height approximately as exp(\( z/2H \)). The density scale height \( H \approx 7 \) km, and so \( \lambda_h \) should increase by a factor of 3 over the 15- to 30-km height range, as observed in Fig. 1B. Thus, the mountain wave propagates without breaking from 15 to 30 km. Above 30 km, the oscillations in Fig. 1B abruptly attenuate and disappear. There are two plausible reasons for this. Because winds weaken rapidly above this level (Fig. 1D), \( \lambda_h \) decreases according to Eq. 1, and the wave soon becomes too short for CRISTA to resolve. Additionally, however, the wave is likely to break above 30 km. Mountain waves advect heavy air over lighter air and overturn when amplitudes exceed the breaking threshold

\[ \hat{T} \geq \hat{T}_{\text{break}} \approx \frac{dT}{dz} \left[ \frac{\Gamma_a}{U} + \hat{T}_{\text{break}} \right] \]

(2)

where \( \Gamma_a \) is the adiabatic lapse rate (9.8 K km\(^{-1}\)) and \( \hat{T} \) is background temperature (19). At 15 to 30 km, the mountain wave-breaking amplitude \( \hat{T}_{\text{break}} \approx 10 \) K. This value is consistent with the observed nondissipative growth in \( \hat{T} \) from 15 to 30 km, because \( \hat{T} < \hat{T}_{\text{break}} \) at these heights. At 30 km, however, \( \hat{T} = 7 \) K (Fig. 1B) and is probably larger in reality, given some observational smearing. Indeed, retrieval calculations indicate that, for a wave of \( \lambda_h = 6.5 \) km and \( \lambda_h = 400 \) km, the amplitude \( \hat{T} \) measured by CRISTA is only ~60% of its true value (17). Thus, this wave is at or near its breaking amplitude at 30 km. Above 30 km, Eq. 2 shows that \( \hat{T}_{\text{break}} \) decreases because of the weakening winds in Fig. 1D, making wave breaking and amplitude attenuation even more probable and severe.

The influence of a breaking mountain wave on the atmosphere is determined by its vertical flux of horizontal momentum density, the so-
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**Abstract:**

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called Eliassen-Palm (EP) flux, which can be computed from measured wave parameters (20). With the use of Eq. 2 and linear saturation theory (2, 19), wave breaking above 30 km yields large EP flux divergences that imply mean-flow accelerations of up to \(-(50 \text{ to } 200)\) m s\(^{-1}\) day\(^{-1}\). These decelerative forces on the eastward stratospheric flow in Fig. 1D are one to two orders of magnitude larger than zonal-mean climatological estimates of small-scale wave forcing at 40º to 50ºS (21). Such large values imply vigorous local wave breaking, drag, turbulence production, and mixing at 30 to 40 km. The wave is likely to dissipate entirely below 43 km, where a critical level (\(U_c = 0\); Fig. 1D) causes both \(\lambda\) and \(T\) to vanish (Eqs. 1 and 2) and prevents further penetration of mountain wave energy (1).

How common are large mountain wave events? Figure 2 compares \(\lambda\) and \((\lambda)_{\text{theory}}\) at \(z = 25\) km based on \(T'(z)\) data for all seven mission days at 40º to 50ºS over South America. Again, there is very good correlation between theoretical and measured vertical wavelengths. Amplitudes measured on four successive days are plotted in Fig. 3, A to D. Underlying plots show corresponding predictions, with the Naval Research Laboratory Mountain Wave Forecast Model (MWFM) (4, 22) interfaced to assimilated regional winds and temperatures from NASA's Data Assimilation Office (DAO) (Fig. 3, E to G) (23). The MWFM results capture much of the day-to-day variability. Figures 2 and 3 reveal that mountain waves were present in this region of the stratosphere throughout the mission.

These long plane stratospheric mountain waves resemble those observed by aircraft over the Scandinavian Ridge (8, 24). Rapid cooling of air by the latter waves produces PSCs, which in turn destroy Arctic ozone (8). Although the Andean mountain waves observed here also produce large temperature decreases (Fig. 3), they are probably too far north to yield temperatures cold enough to form PSCs. Nevertheless, as the edge of the Antarctic vortex moves back and forth across the tip of South America (25), strong turbulence generated by these mountain waves will mix tongues of ozone-depleted vortex air with ozone-rich air from mid-latitudes, thus diluting local ozone concentrations overall. The large EP flux divergences could also modify the subsequent evolution of the vortex (26).

Are stratospheric mountain waves evident elsewhere in the CRISTA data? \(T\) amplitudes over the Northern Hemisphere at 25, 35, and 45 km on 9 November 1994 are plotted in Fig. 4, A to C. Mountain wave amplitudes \(T\) predicted by MWFM on this day are shown in Fig. 4, D and F. Although the peak amplitudes differ (27), observations and model results show very similar geographical distributions. Most notable is a broad region of enhanced amplitudes over central Eurasia. The activity is produced by mountain waves from the Altai Mountains,
Fig. 3. Amplitudes over southern South America from 6 to 9 November 1994 (A to D, respectively), plotted as color pixels at the profile location. Raw values were scaled according to their vertical wavelengths to correct for observational smearing (17). Beneath each panel is a corresponding MWFM prediction based on DAO assimilated winds and temperatures at 12:00 UT for these dates (E to H). Modeled waves with $\lambda_z < 5$ km were not plotted because CRISTA is not sensitive to them. The color scales are linear. Pixels are plotted in order of ascending amplitude and are sized to maximize coverage and visibility while minimizing pixel overlap.

Fig. 4. (A to C) Polar orthographic maps of amplitudes $\hat{T}$ from CRISTA profiles over the Northern Hemisphere on 9 November 1994 at 25, 35, and 45 km, respectively. Values were corrected for observational smearing as in Fig. 3. (D and F) Corresponding MWFM amplitudes $\hat{T}$ at 25 and 45 km, respectively, based on DAO assimilated winds and temperatures at 12:00 UT on this date. As in Fig. 3, modeled waves with $\lambda_z < 5$ km were omitted. (E) Topographic elevation in the Northern Hemisphere. All color scales are linear. Pixels are plotted in order of ascending amplitude.
Crista measured gravity waves from sources in stratospheric winds can modulate the ground gravity-wave activity, due to the temperature structure of the winter polar stratosphere (13, 29), although longitudinal asymmetries in stratospheric winds can modulate the effect substantially (30). This indicates that Crista measured gravity waves from sources other than mountains as well.

Enhancement of Interdecadal Climate Variability in the Sahel by Vegetation Interaction

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The role of naturally varying vegetation in influencing the climate variability in the West African Sahel is explored in a coupled atmosphere–land-vegetation model. The Sahel rainfall variability is influenced by sea-surface temperature variations in the oceans. Land-surface feedback is found to increase this variability both on interannual and interdecadal time scales. Interactive vegetation enhances the interdecadal variation substantially but can reduce year-to-year variability because of a phase lag introduced by the relatively slow vegetation adjustment time. Variations in vegetation accompany the changes in rainfall, in particular the multidecadal drying trend from the 1950s to the 1980s.

The rainfall over the West African Sahel region (J) shows a multidecadal drying trend from the 1950s to the 1980s and early 1990s, as well as strong interannual variability (Fig. 1A). Causes proposed to explain this dramatic trend include global sea surface temperature (SST) variations (2–5) and land use change, that is, the desertification process (6, 7). Because vegetation distribution tends to be controlled largely by climate (8, 9), and surface property changes can affect climate by modifying the atmospheric energy and water budget (10–13), it is reasonable to propose that dynamic vegetation-climate interaction might influence decadal climate variability substantially in a climatically sensitive zone such as the Sahel. We tested this

References and Notes