Tonga eruption triggered waves propagating globally from surface to edge of space

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Abstract

The January 2022 Hunga Tonga–Hunga Ha‘apai eruption was one of the most explosive volcanic events observed in the modern era[1,2], producing a vertical plume which peaked more than 50km above the Earth. The initial explosion and subsequent plume triggered atmospheric waves which propagated around the world multiple times. Here, we combine a comprehensive set of satellite and ground-based observations to analyse and quantify this wave response, from surface to ionosphere. A broad spectrum of waves was triggered by the initial explosion, including Lamb waves[3,4] propagating at 318.2±6 $\mathrm{ms^{-1}}$ at surface level and between 308±5 to 319±4 $\mathrm{ms^{-1}}$ in the stratosphere, and fast gravity waves[5] propagating at 238±3 to 269±3 $\mathrm{ms^{-1}}$ in the stratosphere. Atmospheric gravity waves at sub-ionospheric heights have not previously been observed propagating either at this speed or over the whole Earth from a single identifiable source[6,7]. Latent heat release from water and hot ash in the plume remained the most significant individual gravity wave source at any location for the next 12 hours, producing circular wavefronts visible across the Pacific basin in satellite gravity wave observations. A single source dominating such a large region is also unique in the observational record. The Hunga Tonga eruption represents a key natural experiment in how the atmosphere responds to a sudden point-source-driven state change, which will be of significant use for improving atmospheric weather and climate models.

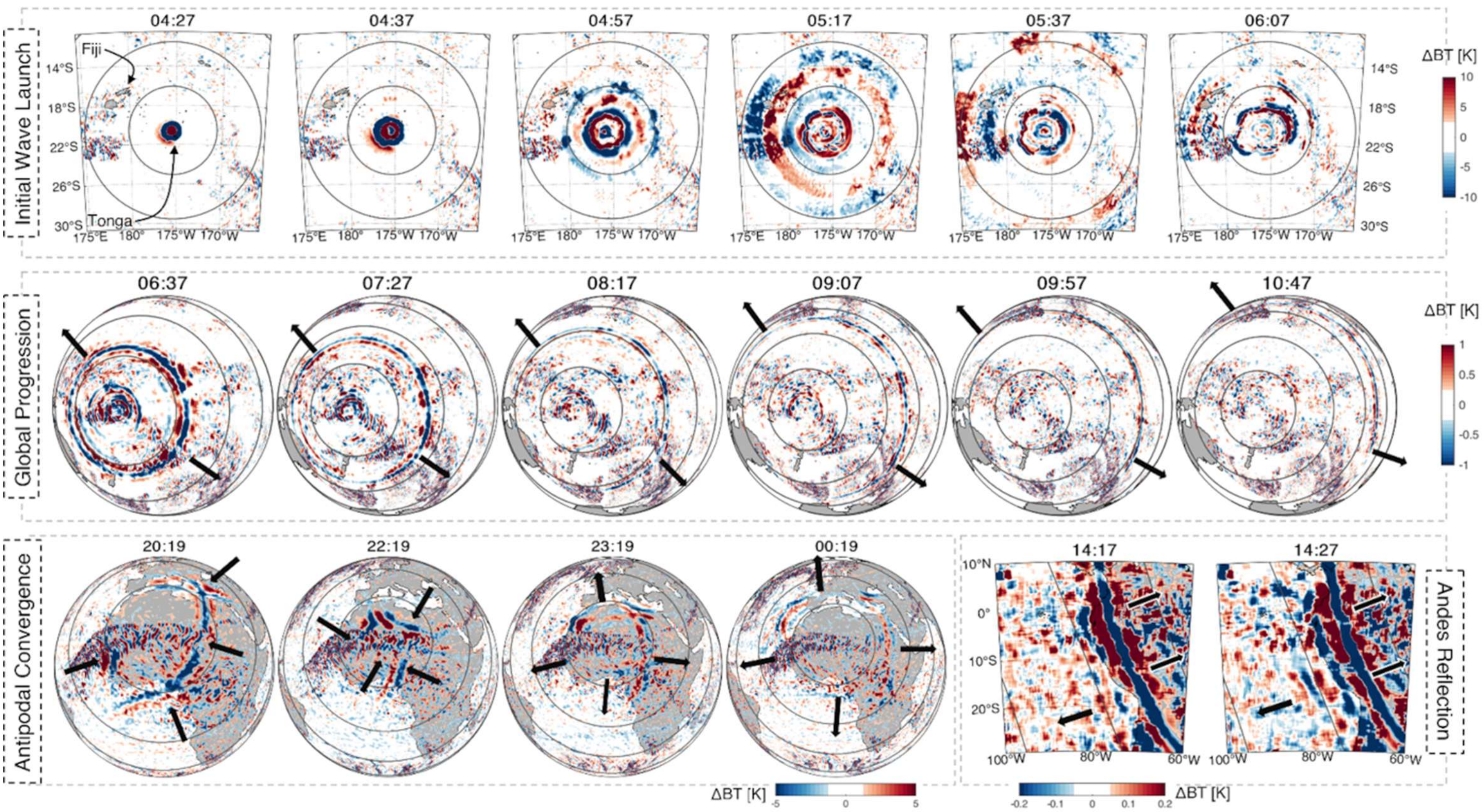
On the 15th of January 2022, the Hunga Tonga–Hunga Ha‘apai submarine volcano (20.54°S, 175.38°W, hereafter ‘Hunga Tonga’) erupted, producing a vertical plume >30 km tall with overshooting tops above 55 km, a record in the satellite era[8] and likely longer[2]. From surface-pressure data, we estimate a single-event energy release from the initial explosion of between 10-28 EJ, likely larger than the 1991 Mt Pinatubo eruption (~10 EJ[2]), and possibly comparable to Krakatoa in 1883 (~30 EJ[2]) (see Methods and Extended Data Figures 1a,b).

Large explosions such as volcanoes and nuclear tests are theoretically understood to produce atmospheric waves[9,10] across a range of length and frequency scales. At horizontally-short wavelengths, these include external Lamb waves[3,4,11] , acoustic waves[10] and internal gravity waves[12]. In addition to explosion-generated waves, volcanoes can also act as a sustained wave source after the initial eruption via updrafts and heating associated with plume convection[13,14].

In practice, observations of such waves at non-acoustic frequencies after volcanic eruptions are rare. Krakatoa[4] and Pinatubo[15], amongst others, produced strong Lamb waves visible in surface pressure. Internal waves in the boundary layer have been inferred from seismography, barometry and infrasound for eruptions including El Chichon[13] (1982), Pinatubo[13] and Okmok[14] (2008). In the free atmosphere, local gravity wave activity associated with plume convection has been seen in mesospheric nightglow over the La Soufriere (2021) and Calbuco[12] (2015) eruptions and in local cloud over eruptions including Cumbre Vieja (2021). Re-examination of 1990s Advanced Very High Resolution Radiometer data also shows waves in cloud above Pinatubo (Extended Data Figure 2). Finally, an electron-density ionospheric wave response is usually observed[16,17], with the response magnitude proposed as a metric of volcano explosive power[18].

There is however no direct observational evidence for long-distance propagation in the free neutral atmosphere of either Lamb or gravity waves triggered by volcanoes. Pre-2000s

*Figure 1: **Initial Lamb wave propagation in the troposphere**: Brightness temperature changes observed by (top two rows) GOES, (bottom left) Meteosat Spinning Enhanced Visible and InfraRed Imager (SEVIRI) and (bottom right) GOES-EAST. Range rings indicate distance from Hunga Tonga in (top row) 500km and (lower rows) 2000km steps. To reduce noise from weather systems, global and antipodal panels have been processed with a 200km-radius Wiener filter, and Andes panels with a 400km boxcar and 72-km-radius Wiener filter. Black arrows indicate approximate wave location and propagation direction. All times UTC.*



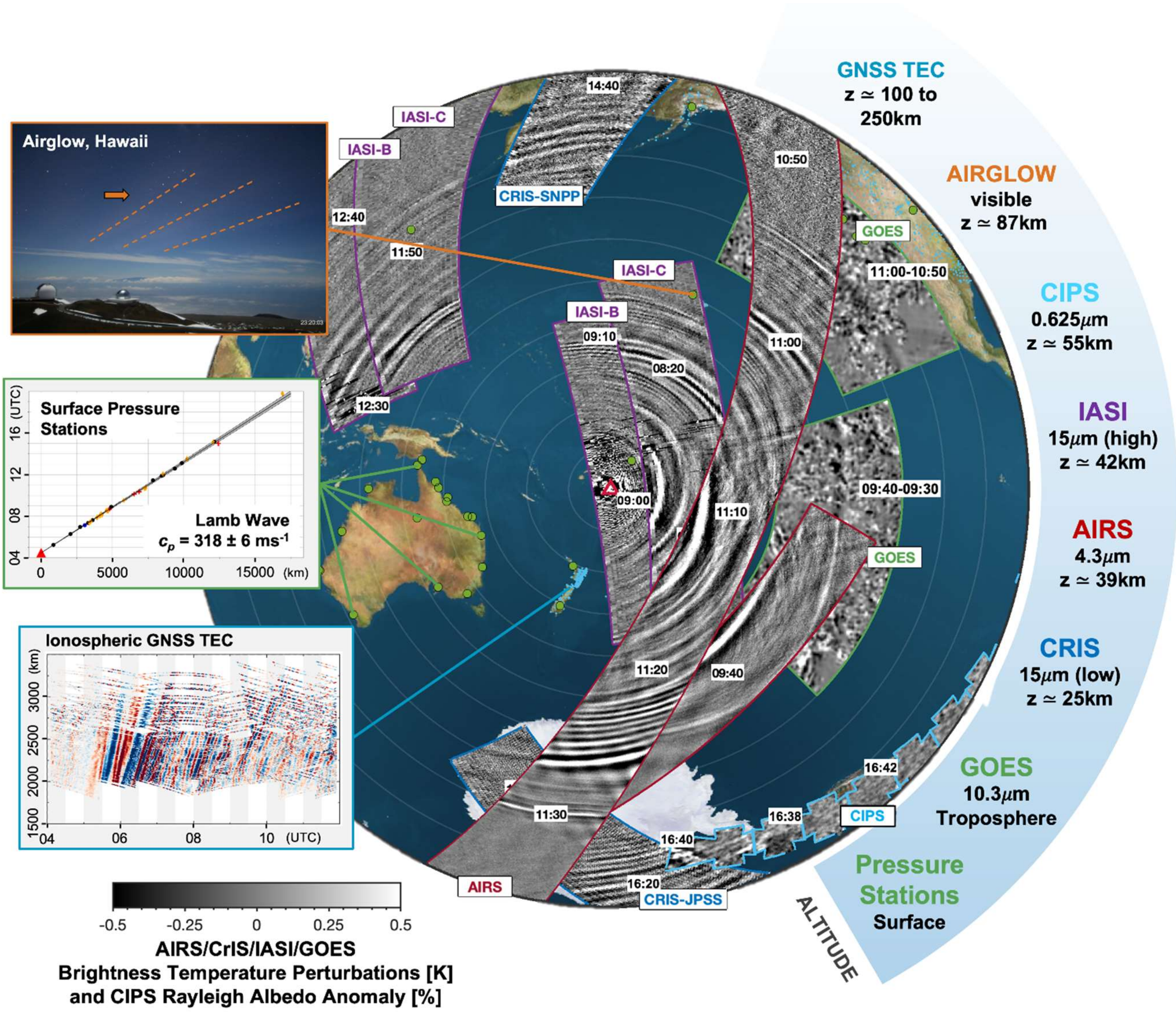


Figure 2: ***Initial gravity and Lamb wave propagation at all heights:*** *Combined measurements of the initial wave release from multiple platforms, listed with their approximate altitudes at right and at times as indicated by overlaid text labels. Inset panels showing pressure (green outline) and TEC (blue outline) distance/time series are reproduced as Extended Data Figures 1d and 3 respectively. Note that AIRS, CrIS and IASI all measure the same three stratospheric altitude channels, but only one is used here from each instrument to show all levels while maintaining visual clarity; due to the long vertical wavelengths of the observed waves, all three levels are near-identical. Airglow inser shows a northward view containing the Lamb wavefront at 09:20 UTC, ~30 minutes after the wave passed overhead.*

satellite observations had insufficient resolution and coverage to measure such waves, and no event since[6] has produced a wave response similar to that identified within hours[19] of Hunga Tonga. This eruption thus represents an opportunity to quantify the wave response to a point-source disruption at a scale and comprehensiveness unique in the observational record.

Eruption and Immediate Wave Response

Figures 1 and 2 show the propagation of Lamb and gravity waves triggered by the initial eruption on the 15th of January, Figure 1 as height-integrated data from the Geostationary

Operational Environmental Satellite (GOES) and MeteoSat platforms and Figure 2 as height-resolved measurements from multiple instrument types in addition to GOES.

The eruption became visible just after 04:00 UTC as a plume which reached a width of 200km and height of >30km within 30 minutes[8]. 20-30 minutes after the plume began rising, a shockwave became visible in ten-minute-resolution near-infrared geostationary imagery. Back-projection from surface pressure data shows that the trigger source occurred at 04:28±2 UTC, with the leading wavefront propagating away at a near-surface phase speed of 318.2±6 ms^{-1} (Figure 2, Extended Data Figure 1c,d, Supplementary Figure 1). Based on the high phase speed, large amplitude and non-dispersive nature of the signal we identify this as a Lamb wave, i.e. a mixed packet of waves with non-dispersive wavelengths and periods travelling at the same speed. This speed is consistent with the Lamb wave produced by Krakatoa, estimated[20] to have propagated at 318.8±3 ms^{-1}.

The Hunga Tonga Lamb wave propagated around the globe, passing through the antipodal point in Algeria 18.1 hours (±7.5 minutes) after the eruption (Figure 1). By this time, the wavefront had deformed due to atmospheric and surface processes, and passed through the antipode as four distinct wavefronts. Over following days, it was tracked propagating at least three times[21] around the Earth. We also see a faint signal in GOES data consistent with the wave being partially reflected from the Andes on its first transit (Figure 1), and evidence of the wave being slowed over South America (Supplementary Figure 2).

Using radiance data from the Advanced Infrared Sounder (AIRS), Cross-track Infrared Sounder (CrIS) and Infrared Atmospheric Sounding Interferometer (IASI) polar-orbiting thermal infrared (IR) sounders (specifically, 4.3µm data sensitive to altitudes ~39 km±5 km and 15 µm data sensitive to the both ~25±5km and ~42±5km altitude levels separately, Figure 2), we see the Lamb wave as a high-amplitude monochromatic pulse with a phase speed of between 308±5 and 319±4 ms^{-1} depending on location. We also observe it as a pulse just above the noise floor of Cloud Imaging and Particle Size (CIPS) Rayleigh albedo data 12 300km away from and 10.75 hours after the eruption (~55±5km altitude, phase speed 316-319 ms^{-1}, Extended Data Figure 4a), and as phase fronts in hydroxyl airglow over Hawai'i, 4960 km away from and 4.3 hours after (~87±4km altitude, phase speed 318 ms^{-1}).

The observed Lamb wave phase fronts are uniform in height and phase speed to within the error range of each instrument from the surface to at least the upper mesosphere/lower thermosphere. The energy density of a Lamb wave is theoretically expected[22] to decay exponentially with height, and the observed phase speed is consistent with a vertical mean of sound speed weighted according to this energy distribution (see Methods). We observe a slightly different speed for propagation in different directions across the Earth (e.g. at Broome, Australia, we measure 319 ms^{-1} for the westward-travelling wave and 316 ms^{-1} for the eastward, Extended Data Figure 1e), and the asymmetric perturbations we observe are consistent in sign with such a shift due to background winds.

Following the Lamb wave, we observe a series of slower waves with continually varying speeds and horizontal wavelengths (λ_h) that we identify as a dispersive packet of fast internal gravity waves (Figure 2). These have phase speeds of 240-270 ms^{-1}, varying with local λ_h. The leading phase front has the largest amplitude and longest λ_h, with a brightness temperature (BT) amplitude of 0.74 K and λ_h of 380 km here falling to 0.15 K and 100 km across the packet width. This packet is observed to extend ~2000 km and eight phase cycles

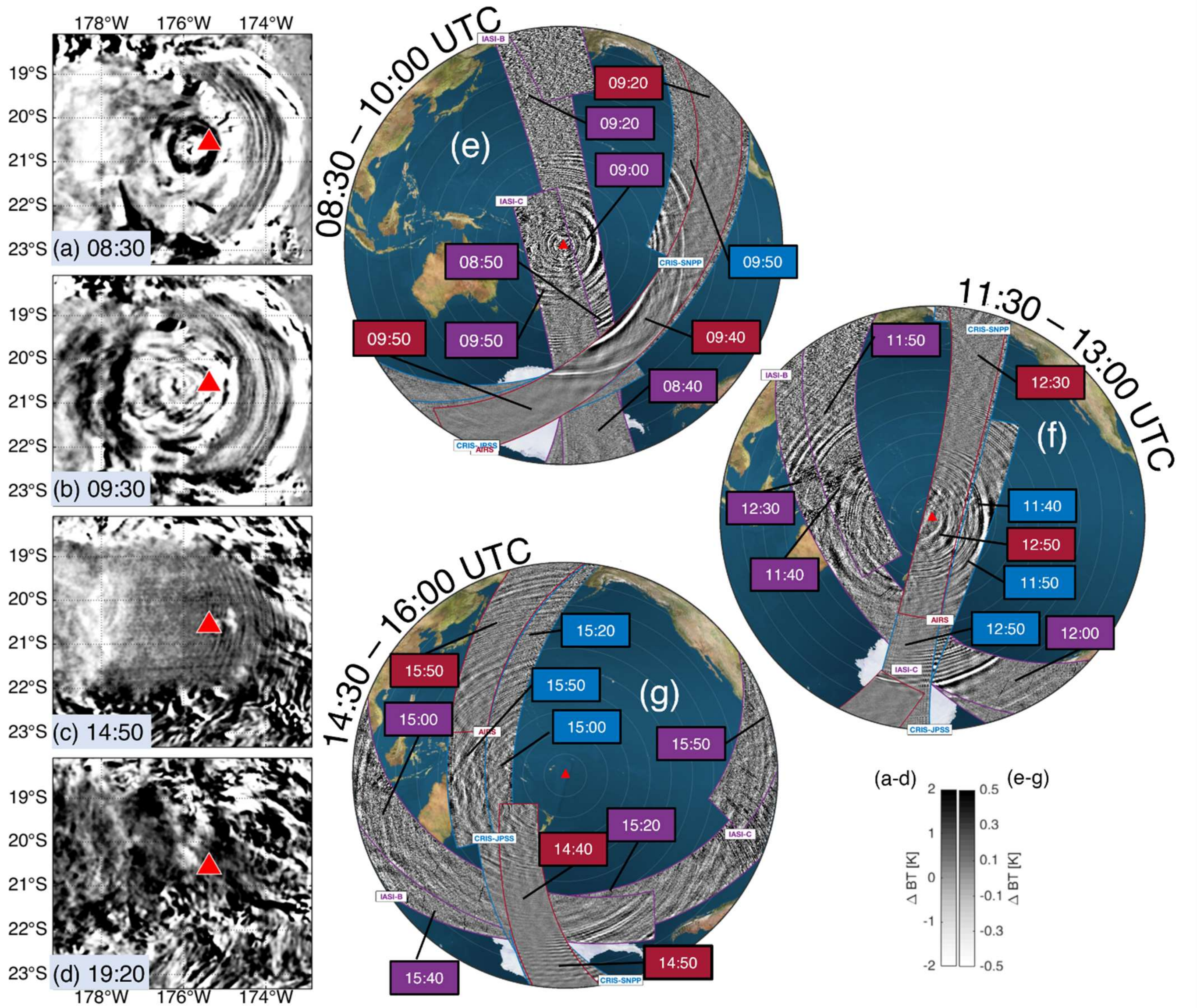


*Figure 3: **Post-eruption wave activity**: (a-d) in and around the volcanic plume as observed by GOES and (e-g) over the entire Pacific basin as observed by AIRS, CrIS and IASI. For (e-g,) coloured labels indicate individual satellite overpass times for context, with AIRS labelled in red, CrIS in blue and IASI in purple. Note that the colour scales in panels (a) and (b) saturate significantly, and values extend to +-8K.*

across the South Pacific ~7 hours after generation (Extended Data Figure 5). We observe the packet over multiple orbits of AIRS, CrIS, and IASI across the globe, in CIPS over Antarctica, and in airglow (~85km altitude, depth ~8km) above Hawai'i. Vertical wavelength (λ_z) is poorly defined but very deep: no phase difference is seen between AIRS observations at 25 and 42 km altitude, and calculations based on observed speed and λ_h imply $\lambda_z >> 110$ km, i.e. greater than the depth of the homosphere. These phase speeds are consistent with vertically-propagating gravity waves travelling at speeds close to, but very slightly less than, the theoretical maximum speeds achievable prior to total internal reflection (See Methods and Extended Data Figure 6) and with the same temporal origin and source as the Lamb wave.

This leading gravity wave packet passes through the antipode at times between ~00:30 and 02:30 UTC, i.e. 20-22 hours after the eruption (Extended Data Figures 7a-c), with the broad time window determined by separation of different λ_h components with time. Gravity waves remaining coherent and expanding over the whole globe from a single source of any kind are unprecedented in the observational record[6]. On their return journey from the antipode, the waves become difficult to distinguish in our intermittent low-Earth orbit satellite snapshots from those produced both later by Hunga Tonga and by other sources, and consequently we cannot track them to their extinction.

The gap between the initial Lamb wave and subsequent gravity wave grows with time. This is consistent with a theoretically-predicted forbidden phase speed range between external Lamb wave and internal gravity wave limits imposed by total internal reflection (Extended Data Figure 3). Two low-amplitude wavefronts are present in the gap; these propagate with the same speed as the leading Lamb wavefront, but trace back to different origin times (Figure 2 and Extended Data Figure 5b). We therefore identify these as Lamb waves triggered by subsequent smaller explosions which were also observed in local surface pressure (Extended Data Figure 8).

Ionospheric data (Figure 2 and Extended Data Figure 3) show key differences from the lower atmosphere. Over New Zealand, we see three large travelling ionospheric disturbances (TIDs), with phase speeds, λ_h and amplitudes of (1) 667 ms^{-1}, 1000 km, 0.2 TEC Units (TECu); (2) 414 ms^{-1}, 700 km, 0.4 TECu and (3) 343 ms^{-1}, 400 km and >1 TECu respectively. They are consistent in speed and direction with a Hunga Tongan source between 04:15 and 05:00, but do not share the arrival time, phase speed or λ_h of the Lamb wave in other atmospheric layers. Therefore, we do not identify these TIDs as the Lamb wave. However, a strong and brief TEC modulation, spiking at an amplitude of >0.6 TECu, is seen at 6.15am consistent with the expected arrival time and brief period of the Lamb wave.

We do not see TID 1 over North America, but do see a signal consistent with TID 2 and another TID (4) with phase speed ~311 m/s which is also consistent with TIDs measured over New Zealand. We again see a strong TEC modulation at the expected Lamb wave arrival time.

The properties of TIDs 1 and 2 are inconsistent with slant path gravity waves propagating from Hunga Tonga, but could have reached the observed sites by indirect paths, e.g. by vertically propagating as acoustic or gravity waves above the volcano then travelling at high horizontal speeds through the ionosphere. The properties of TIDs 3 and 4 are consistent with the wave activity generated over Hunga Tonga in the hours after the primary eruption.

Sustained Post-Eruption Wave Generation

After the initial trigger, sustained gravity wave generation is seen in the clouds above Hunga Tonga and radiating outwards across the Pacific basin. While smaller in amplitude and slower in phase speed than those from the initial eruption, these waves are also highly anomalous relative to past gravity wave observations.

Figure 3 shows BT measurements from (a-d) the GOES 10.3μm channel over the Hunga Tonga area and (e-g) the AIRS, CrIS and IASI 4.3μm stratospheric channels over the Pacific basin for selected times.

In GOES observations of the eruption cloud top (Figure 3a-c, Supplementary Figure 3), arced features consistent in morphology and temporal progression with propagating concentric gravity wave phase fronts are visible. λ_h ranges from the 8km resolution limit of the data to 65km, and BT amplitude from 0.5-8K. These measured properties are very similar to those of gravity waves generated near the convective centres of hurricanes.

The apparent centre of these waves is slightly west of Hunga Tonga. This is consistent with refraction of the wave field by the prevailing easterly winds. The waves are remarkably consistent in concentric shape over several hours, suggesting a powerful and relatively persistent pulsing source for wave generation. The source may be pulses of convection within the plume above the volcano. The waves weaken in amplitude over time, particularly after 15:00UTC, but are visible until at least 19:20 UTC (Figure 3d). They are not found on subsequent days. These results suggest that the volcano may have created a sustained source of convectively-generated waves for nearly fifteen hours after the initial eruption

Stratospheric AIRs, CrIS and IASI observations (Figure 3e-g, Extended Data Figure 7d-o) show wave activity across a range of spatial, frequency and amplitude scales throughout the Pacific basin, all centred on Hunga Tonga. Tracking individual phase fronts is challenging as these data are near-instantaneous at any given location, but conservatively the distribution must include a large fraction of waves with phase speeds >100 ms^{-1}. For example, small-scale continuous wavefronts centred on Hunga Tonga are clearly visible near Japan before 16:00 in Figure 3g and, even if emitted at the earliest possible time of 04.28 UTC, must have phase speeds ~200 ms^{-1} to have travelled this far. Unlike more typical observed waves, these waves can therefore propagate with little apparent influence from global wind patterns due to their unusually large phase speeds. Such fast speeds reduce normal dissipation effects, allowing the waves to propagate vast distances and affect much higher altitudes than typical gravity waves.

These waves dominate the stratospheric gravity wave spectrum over a radius >9000km for >12 hours (Extended Data Figure 7d-o). This is exceptional for a single source, and unique in our observational record[6,7]. Orographic wave sources often persist for longer, but are spatially localised; while some waves in the southern polar jet may have propagated downstream[23,24] or laterally [6,25] from orographic sources, the area they affect is an order of magnitude smaller than here and the waves themselves highly intermittent. Waves from non-orographic sources such as tropical convection and extreme events such as hurricanes, meanwhile, typically become indistinguishable from background within 2000-3500 km[26–27].

How were the waves generated?

Although we cannot directly observe the generation of the waves due to insufficient temporal resolution (for the initial explosion) and ash plume blocking effects (for both the initial explosion and subsequent wave generation), the observed wave properties and context allow us to infer likely mechanisms by which they were generated.

The strong initial response is likely due to the eruption's shallow submarine context and large explosive power. As the volcanic vent was only tens to hundreds of metres below water[28] the seawater did not suppress the blast but was instead flash-boiled[29] and propelled into the stratosphere. Here it condensed, releasing latent heat near-instantaneously across a depth of tens of kilometres. This strong and short-lived forcing would produce vertically-deep waves across a broad spectrum, consistent with observations. This mechanism is also consistent with significant and large IASI-observed increases in stratospheric water vapour (Extended Data Figure 9), and H_2SO_4 in the plume relative to what would be expected for an eruption of this size, which is in turn consistent with sulfuric acid forming in situ due to insufficient volcanogenic SO_2 release and the time available to produce H_2SO_4.

Subsequent wave generation is likely due to similar processes as standard convective waves, such as mechanical oscillator effects[30] associated with vertical air motion within the plume or pulsing from the volcanic heat source below. Such forces would produce sufficiently strong perturbations to generate gravity waves visible both in the plume and propagating freely away. Such a mechanism is again consistent with our observations, particularly the similarity in morphology and amplitude of the observed waves to those generated by hurricanes and convective weather systems.

Weather and Climate Forecasting Implications

While in recent years we have been able to routinely characterise gravity waves in observational data, understanding how the observed spectrum at a location arises has been complicated by fundamental problems in distinguishing the source of a wave from the pathway it has taken to the observation[24]. Being able to separate these problems would lead to major advances in simulating and parameterising gravity waves in next-generation weather and climate models.

The Hunga Tonga eruption represents an important natural experiment in this area. The volcano was a clearly-identifiable near-point source, produced gravity waves across a broad range of spatiotemporal and frequency scales, and these waves were observed by a diverse constellation of instruments worldwide. As such, simulating this eruption in atmospheric models, whether as a point convective source or in a dedicated volcanic simulation, could provide major insight into the strengths and deficiencies of models. In addition, comparison of modelled and observed propagation delays for both the Lamb and gravity waves will provide important information quantifying how well current and future models represent atmospheric winds, temperatures and density structures.

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Tables

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Methods

Explosive Energy Estimate from Surface Pressure Data

We estimate the explosive energy associated with the eruption using three separate approaches. All three give a value in the range 10-28 EJ.

1. Waveform based on a nuclear explosion: Posey and Pierce (1971)[33] suggested that the energy yield of an explosion in the atmosphere can be calculated as $E = 13p\sqrt{[r_e \sin(r/r_e)]}H_s(CT)^{3/2}$, where p is the measured pressure anomaly, r the distance from the explosion, r_e the Earth's radius, H_s the atmospheric scale height, c the speed of the wave, and T the time separation between the first and second peaks of the pressure disturbance. From available pressure-station data at distances ranging from 2500-17500 km from Hunga Tonga (Extended Data Figure 1b), this provides an estimate ~20±8 EJ.

2. Waveform based on previous volcanic eruptions: Gorshkov (1960)[34] estimated the explosive energy of a volcanic eruption as $E = \frac{2\pi H_s \sin(\theta)}{\rho c} \int p^2\, dt$, where θ is the distance from the eruption in degrees and ρ the Earth's surface air density, and t is time. This gives an estimate of ~10EJ.

3. Estimated pressure force: assuming the pressure anomaly spreads under an even cloud of area A, then the work done by the pressure impulse over a column of height h_c is $W = pAh_c$. For an area of radius 200 km and pressure change of 5 hPa, this gives a work estimate ~18 EJ.

Estimate of Lamb Wave Phase Speed

We use the approach of Bretherton (1969)[22] and initial-release data from the European Centre for Medium-Range Weather Forecasts' Fifth-Generation Reanalysis (ERA5T) to calculate the expected speed of the Lamb wave. We first compute the local speed of sound as $c_s(z) = 20.05\sqrt{T}$, where z is the altitude and T the local temperature. For a Lamb wave, where energy density decays exponentially with height, energy density is $E(z) = C\, exp(-z/H)$, where C is a constant term which subsequently cancels in our calculation, and H is

$$H = \frac{c_s^2}{(2-\gamma)} g,$$

for a ratio of specific heats γ which we set to 1.4, and acceleration due to gravity g which we set to 9.80665ms^{-1}. We then calculate the phase speed of the Lamb wave as a vertical mean of the speed of sound weighted by energy density, i.e.

$$c_m^2 = \frac{\int [c_s(z)+u(z)]^2 E(z) dz}{\int E(z) dz},$$

where u is the local wind speed.

For ERA5T meteorological output for the 15th of January 2022 at the 04:00 UTC timestep, this gives a phase speed of 313-318 ms^{-1}. Similar results are obtained using the 05:00 UTC timestep. Our calculation omits the contribution of altitudes above 80 km to the energy density calculation as ERA5 data do not extend above this level, but as energy density decreases exponentially with height this contribution should be small.

Gravity Wave Speed Limit Calculation

Linear wave solutions to the Navier-Stokes equations of the form $A\ \exp[(i(kx + mz + \hat{\omega}t)]$ satisfy the dispersion relation [22] of Fritts and Alexander (2003)[5], which is fourth-order in intrinsic frequency $\hat{\omega}$. For higher-frequency waves where $f^2 \ll \hat{\omega}^2$ and simplifying to planar 2D propagation, i.e. $l = 0$, we can rewrite this as a fourth-order equation in intrinsic phase speed $\hat{c} = \hat{\omega}/k$, i.e.

$$\frac{\hat{c}^4}{c_s^2} - \hat{c}^2\left(1 + \frac{1}{4H^2k^2} + \frac{m^2}{k^2}\right) + \frac{N^2}{k^2} = 0.$$

Letting $x = \hat{c}^2$ gives a quadratic form of the equation

$$ax^2 + bx + c = 0$$

where $a = 1/c_s^2$, $b = -(1 + 1/(4H^2k^2) + m^2/k^2)$ and $c = N$^2/k^2, with solution

$$\hat{c}^2 = \frac{-b \pm \sqrt{b^2 - 4a}}{2a}.$$

Allowing vertical wavenumber $m \to 0$ gives the curve $\hat{c}_{max}(k)$, the maximum phase speed for gravity waves before total internal reflection would prevent their vertical propagation. This limit is

$$\hat{c}_{max}^2 = \frac{c_s^2}{2}\left[1 + (4H^2k^2)^{-1} - \sqrt{[1 + 1/(4H^2k^2)]^2 - 4N^2/(c_s^2k^2)}\right]$$

and is shown as a function of horizontal wavelength k^{-1} in Extended Data Figure 6. Our results for the wave properties produced by Hunga Tonga are consistent with previous theoretical work considering normalised full spectra of acoustic and gravity waves[35].

Airglow Imagery Processing

Airglow data have been obtained from the all-night cloud cameras at the Gemini Observatory on Mauna Kea, Hawaii. Images have been converted from original northward and upward viewing camera angles to an overhead latitude-longitude grid by visual identification of multiple bright stars in the image fields-of-view, then a geometric conversion to give the position of each pixel on the sky at the 87km airglow layer we assume to contain the waves. This assumed height layer is based on the colour of the airglow and spectral range of the cameras used at Gemini, which are both consistent with the hydroxyl (OH) airglow layer.

AIRS, CRIS and IASI

We use brightness temperature observations associated with radiances in the 4.3 µm and 15 µm carbon dioxide absorption bands of AIRS, CrIS, IASI-B and IASI-C[31] on the 15th of January. These instruments can directly resolve stratospheric waves with vertical wavelengths ≳15km and horizontal wavelengths ≳30km, and typically provide twice-daily near-global coverage for each instrument in near-real time with an orbit approximately every 90 minutes. Perturbation fields suitable for spectrally and visually analysing wave signatures are produced by subtracting a fourth-order polynomial in the across-track direction from the data, consistent with previous work using these data[6,32].

CIPS

Imagery from the nadir-viewing CIPS instrument is analysed for the presence of deviations from a smooth model background of Rayleigh scattered UV sunlight (265 nm). The model removes the geometrical dependence of the observation and large-scale geophysical variability of the observed albedo. The data are binned to a uniform 7.5x7.5 km grid, allowing for observations down to 15 km horizontal wavelength. The altitude kernel limits sensitivity to vertical wavelengths $\gtrsim$10km, with peak contribution at ~50 km altitude. The satellite is in a sun synchronous polar orbit with an equator crossing currently near noon.

GOES/MeteoSat

We use data from band 13 of GOES-EAST and GOES-WEST, and band 5 of Meteosat-SEVIRI. These instruments image the Earth's disc at a spatial resolution of 2 km and a temporal resolution of 10 minutes (15 minutes for SEVIRI). Raw radiance data have been converted to brightness temperatures based on the centre wavelength of the channels filters, and then differenced between adjacent timesteps to highlight wave structure.

TEC

Total electron content observations were derived from dual-frequency GPS receivers in the New Zealand GeoNet and the NOAA CORS Networks. Satellite to ground GPS signals were processed following the method of Afraimovich at al (2000)[36], and the dTEC values are projected onto an ionospheric shell altitude of 250 km. The dTEC are then analysed to investigate the travelling ionospheric disturbance parameters.

Data Availability

Airglow data are available from from https://www.gemini.edu/sciops/telescopes-and-sites/weather/mauna-kea/cloud-cam/allnightlong.html. They were obtained under a Creative Commons Attribution 4.0 International License issued by the NSF's NoirLab.

AIRS and CrIS data are available from the NASA Goddard Earth Sciences Data and Information Services Center: https://disc.gsfc.nasa.gov/.

CIPS data are available from the Laboratory for Atmospheric and Space Physics at the University of Colorado Boulder: https://lasp.colorado.edu/aim/.

ERA5 data are available from the Climate Data Store, https://cds.climate.copernicus.eu.

GOES data are available from the NOAA Geostationary Satellite Server, https://www.goes.noaa.gov/.

IASI data are available from the IASI Portal, https://iasi.aeris-data.fr/.

Surface Pressure data are included as a Supplementary file to this manuscript.

TEC data are available from https://www.geonet.org.nz/ and https://geodesy.noaa.gov/CORS/

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Code Availability

All software used is either already publicly available, implements equations provided in the Methods section directly, or only plots data.

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Author Contributions

Conceptualisation: Wright, Hoffmann, Osprey
Data curation: Hoffmann, Bouillon, Carsten, Clerbaux, Mitchell, Randall
Formal analysis: Hindley, Alexander, Barlow, Mitchell, Prata, Hoffmann
Funding acquisition: Wright, Clerbaux
Investigation: All
Methodology: Wright, Hindley, Alexander, Barlow, Mitchell, Prata, Hoffmann
Software: Wright, Hindley, Alexander, Barlow, Mitchell, Prata, Hoffmann
Administration: Wright
Visualisation: Wright, Hindley, Alexander, Barlow, Prata
Writing – original draft: Wright
Writing: review/editing: Alexander, Hoffmann, Mitchell, Osprey

Competing Interest Declaration

The authors declare no competing interests.

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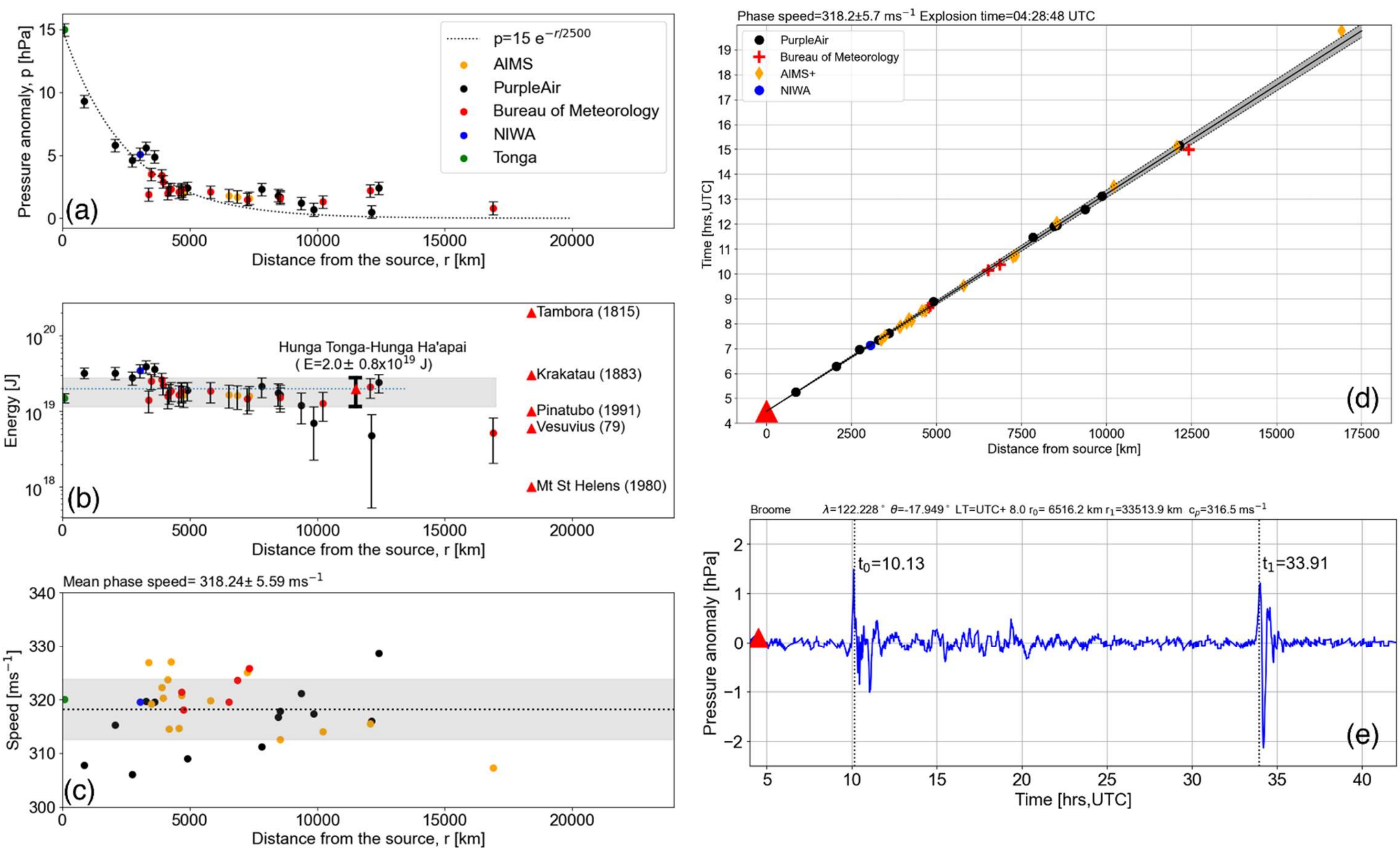
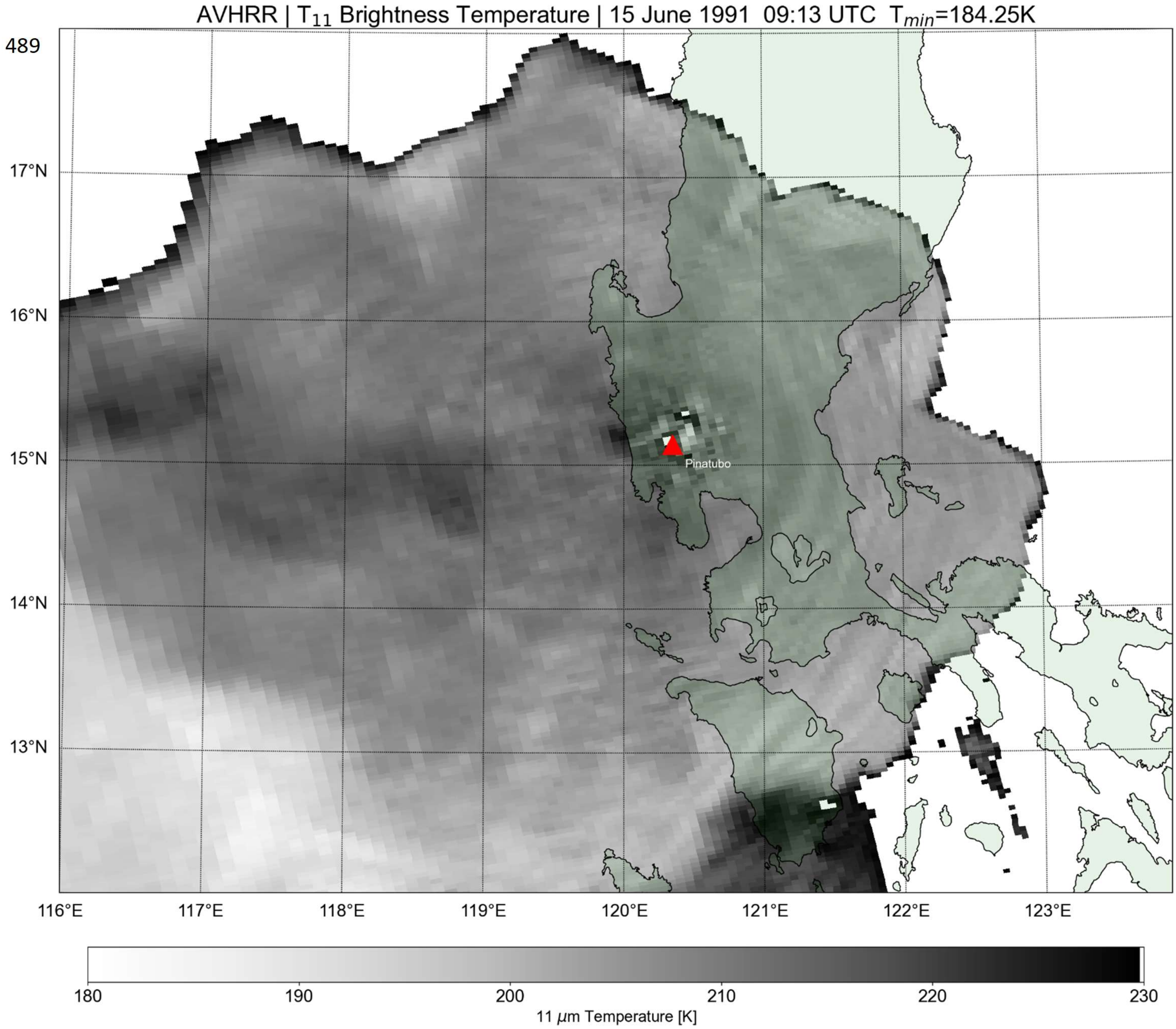
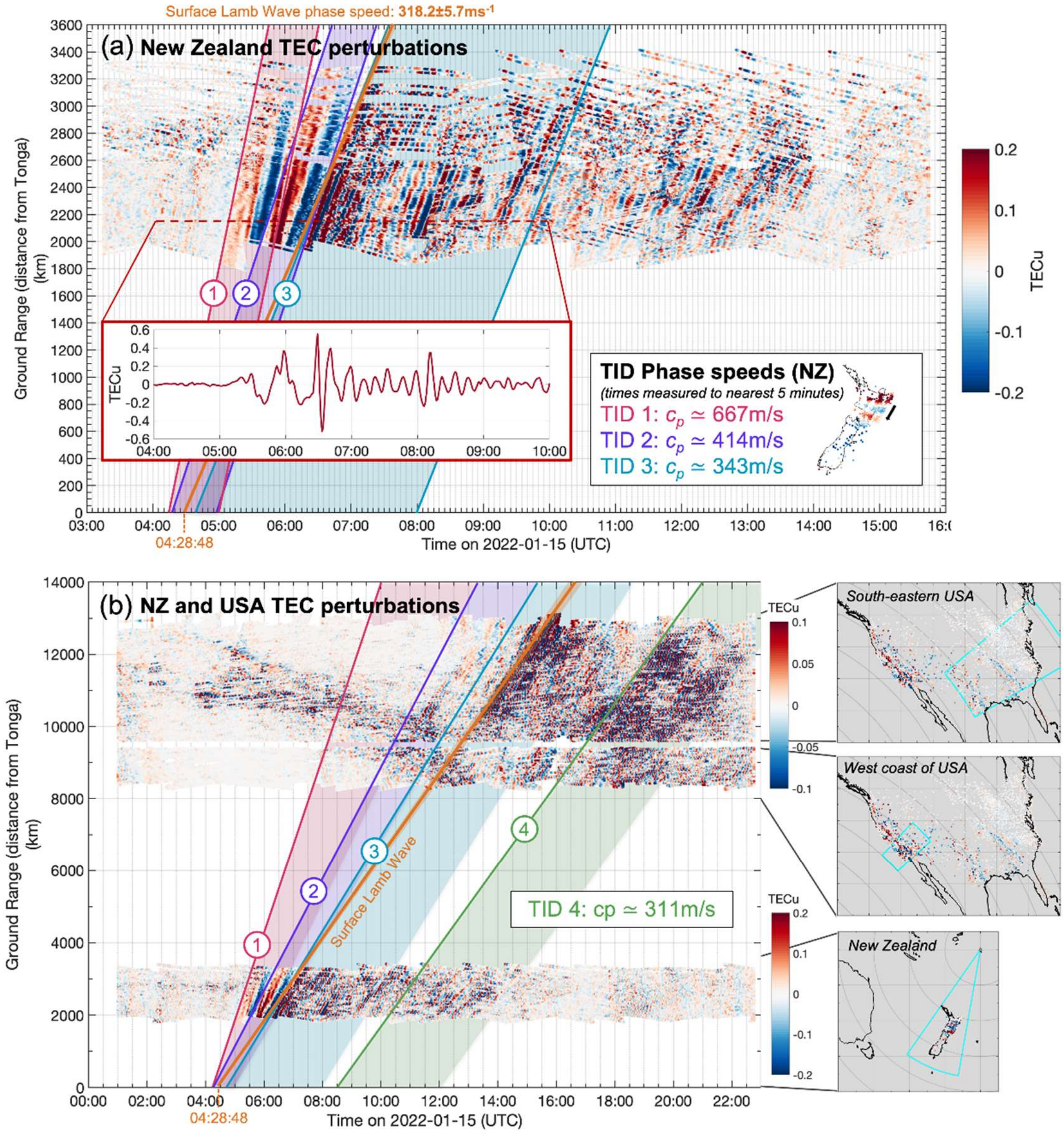
Extended Data



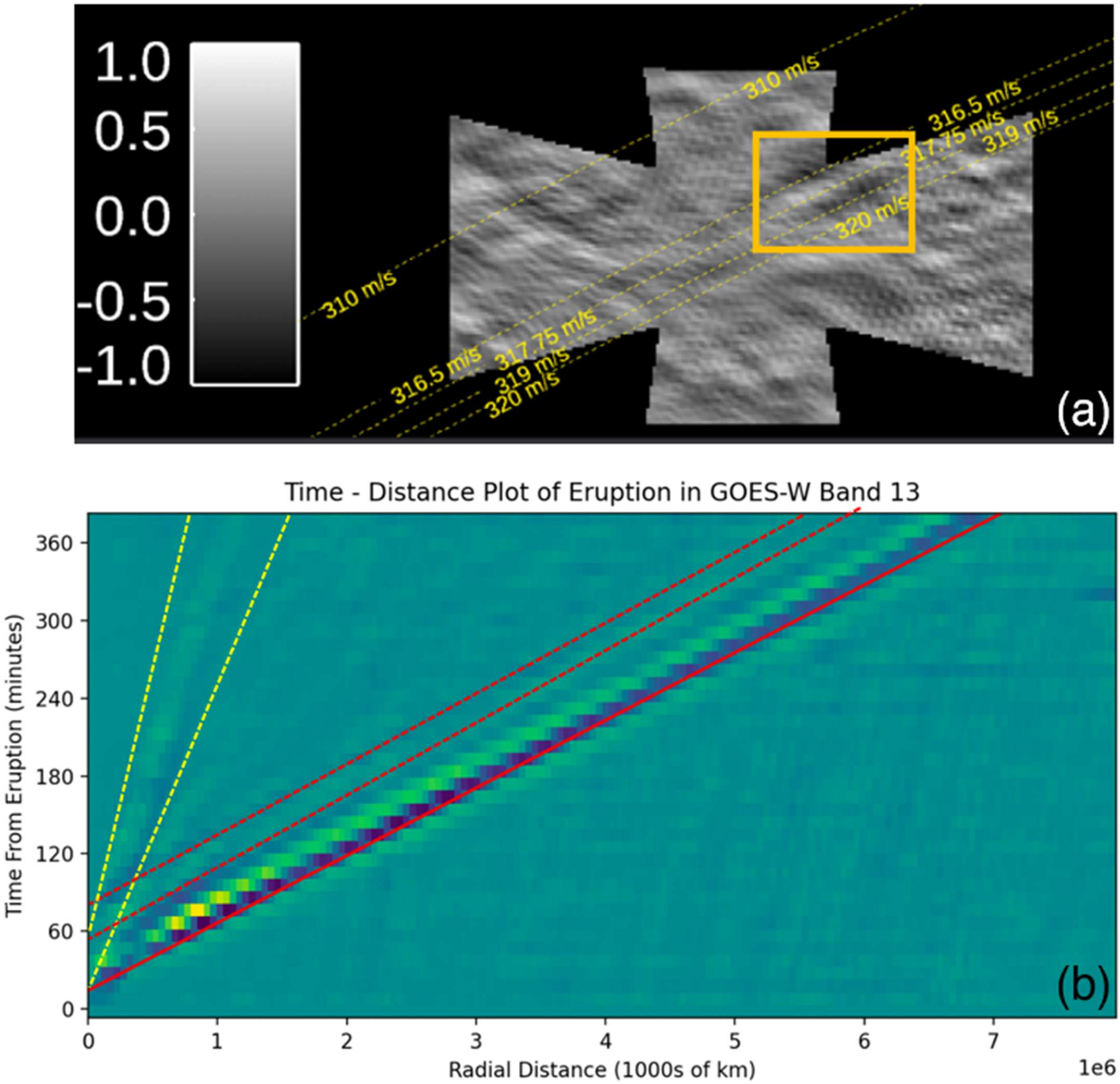
Extended Data Figure 1: (a-d) Estimates of (a) Lamb-wave-induced pressure anomaly, (b) eruption explosive energy, (c) Lamb wave phase speed and (d) time of primary explosion, as computed from surface pressure data. (e) Time series of measured pressure anomaly at Broome, Australia.



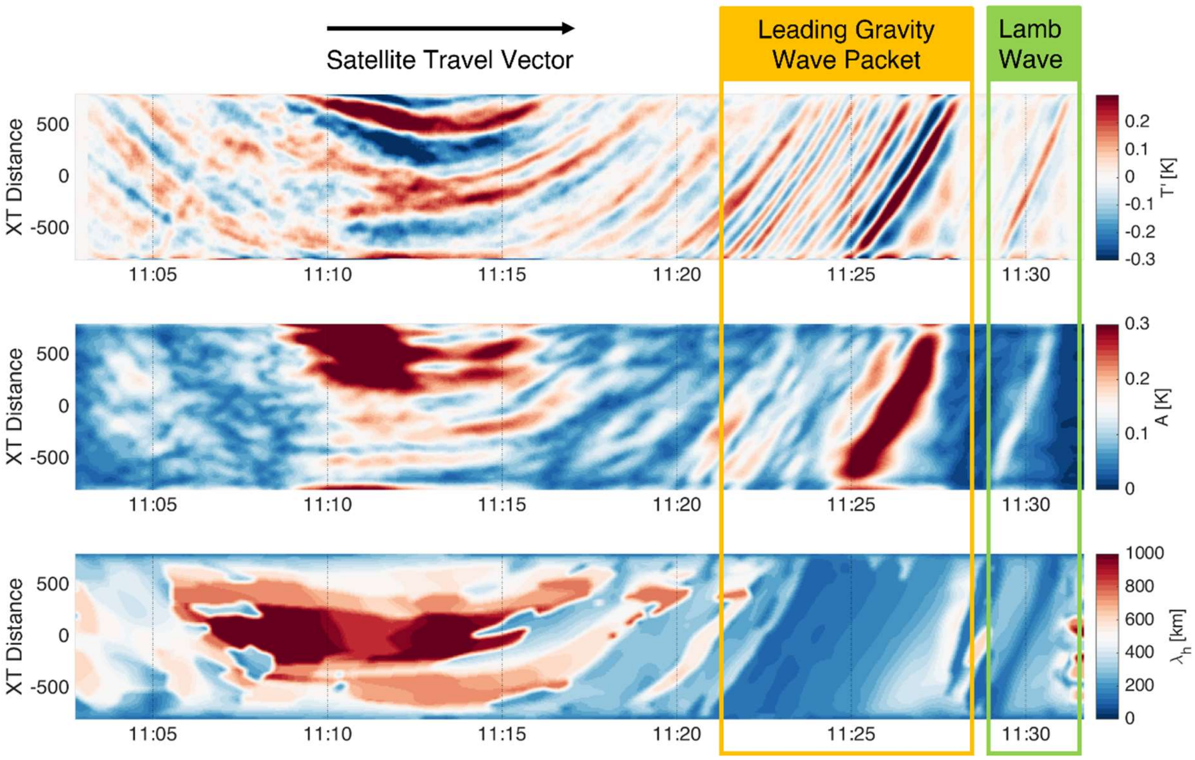
Extended Data Figure 2: Brightness temperature measures over the 1991 Pinatubo eruption plume, as observed by the Advanced Very High Resolution Radiometer.



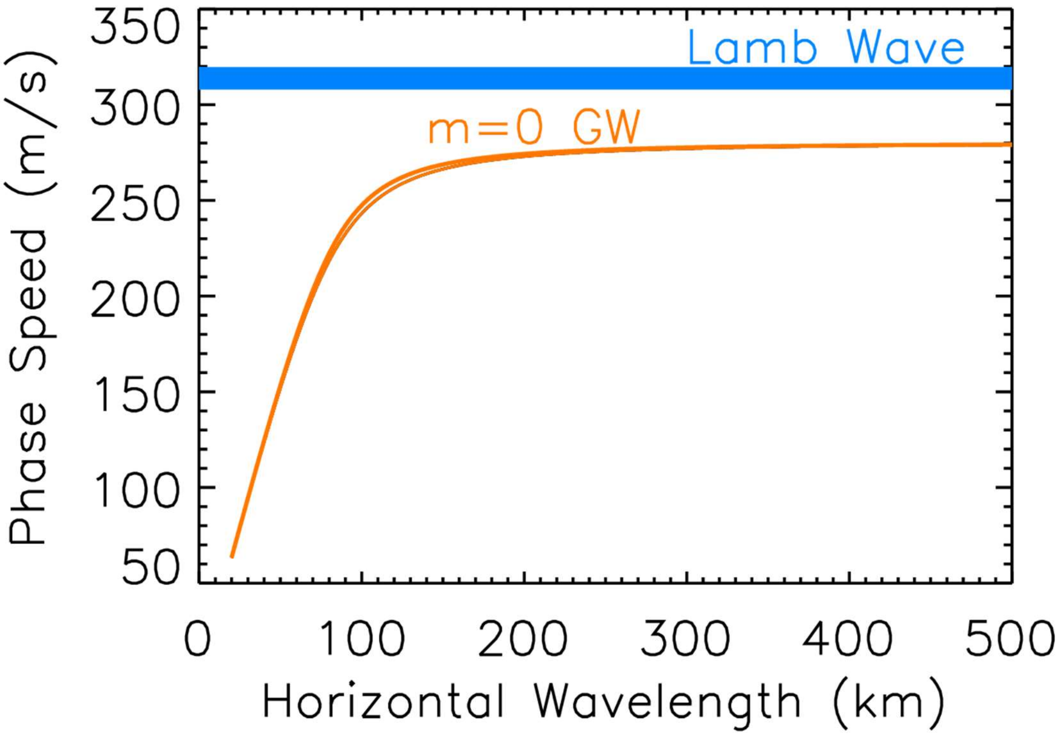
Extended Data Figure 3: Time-distance plots of ionospheric disturbances over New Zealand and the United States, computed from GNSS-TEC data.



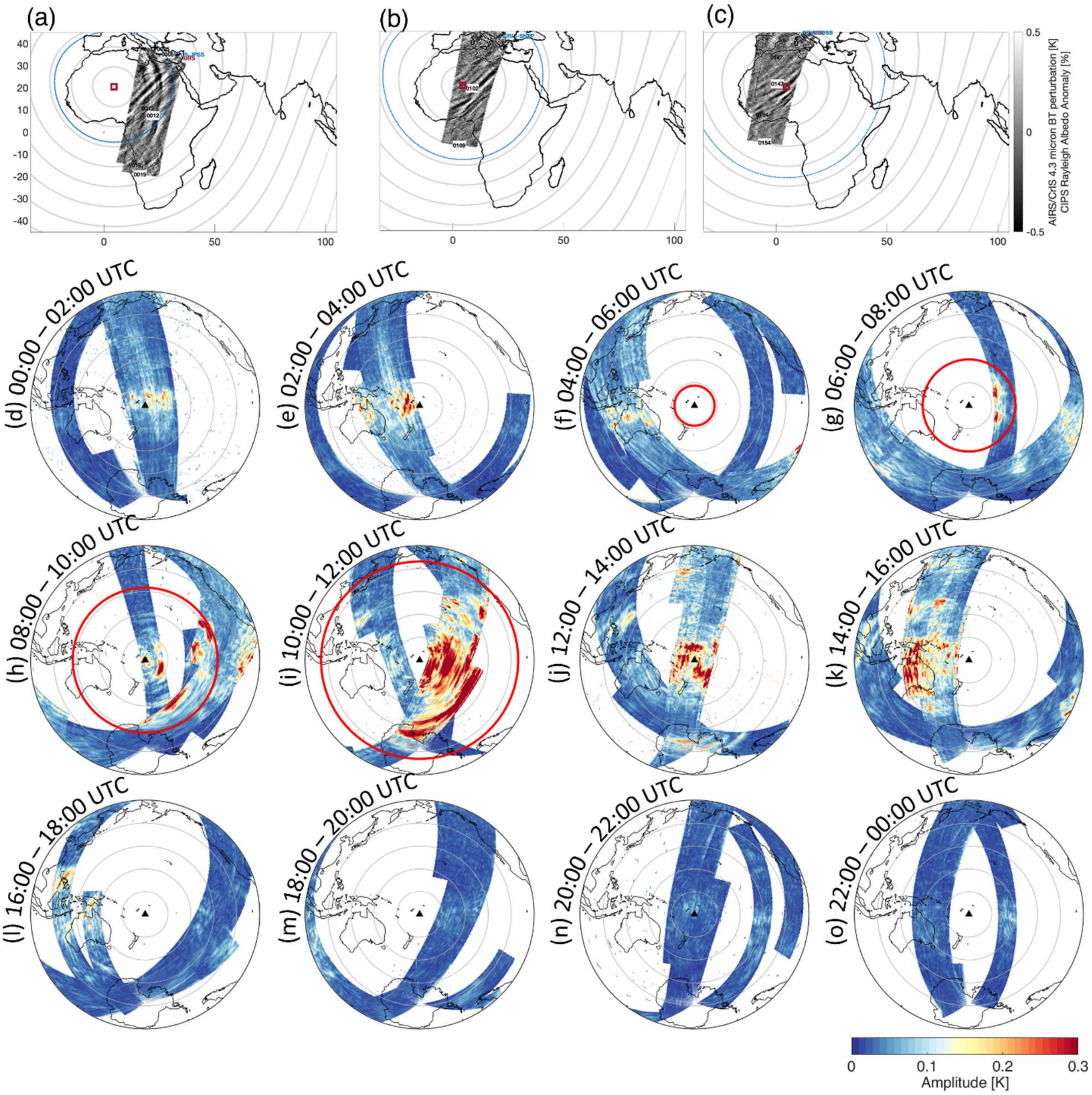
491 Extended Data Figure 4: (a) Lamb wave as observed by CIPS (centred at 24°S 309°E, 12300 from Hunga Tonga, and recorded 10.75 hours after the eruption). In these data, the Lamb wave is extremely close to the instrument noise floor and statistical tests were carried out to confirm that the small signal seen is consistent with the expected speed and wavelength of the Lamb wave. (b) Time-distance spectrum derived from GOES 10um channel, with Hunga Tonga located at the origin. Red solid line identifies the primary Lamb wave, red dashed lines weaker secondary Lamb waves, and yellow dashed lines outline the limits of the dispersive gravity waves in the initially-released packet.



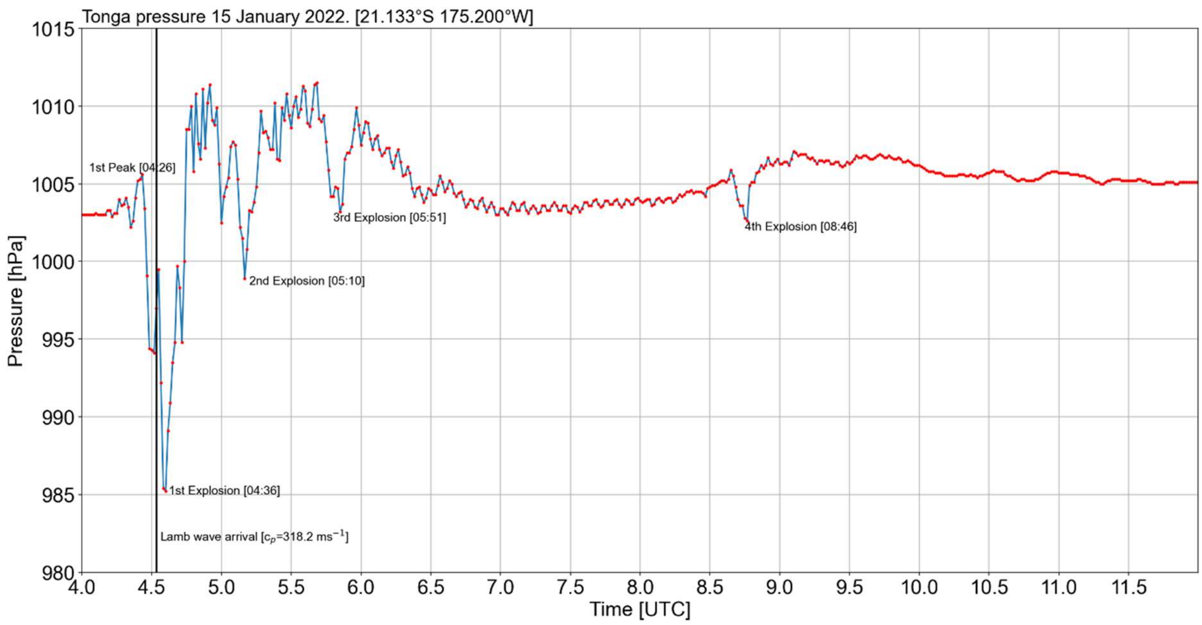
Extended Data Figure 5: 2D S-Transform[37] (2DST) estimates of gravity wave properties measured by AIRS in a descending-node pass over the Pacific Ocean on the 15th of January 2022. (top) temperature perturbations relative to a fourth-order polynomial fit across track. (middle) amplitudes estimated from these perturbations using the 2DST. (bottom) horizontal wavelengths estimated from these perturbations using the 2DST.



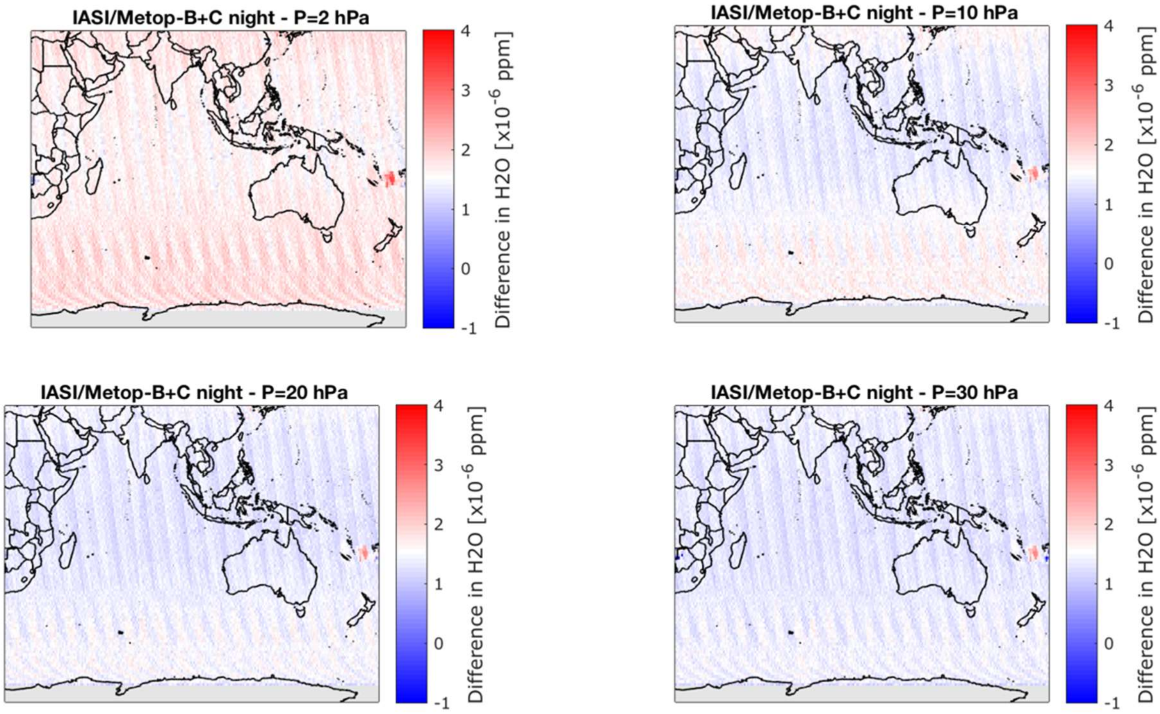
Extended Data Figure 6: Expected maximum speed of a gravity wave packet relative to the observed Lamb wave, as a function of horizontal gravity wave wavelength. Blue line thickness represents the range of Lamb wave propagation speeds that we compute from AIRS, with the fast edge being approximately equal to the speed of the surface pressure signal. Orange lines represent the fast limit of gravity wave phase speeds versus horizontal wavelength, which is in the limit that the vertical wavenumber—>0. This has been calculated using the upper and lower Lamb wave speeds as the sound speed for this calculation, shown as two closely-overlaid orange lines.



Extended Data Figure 7: (a-c) transit of the leading gravity wave packet over the antipode in CrIS and AIRS 4.3 μm data (d-o) GW amplitudes over Pacific computed from AIRS, IASI and CrIS 4.3 μm data using the 2DST[37].



Extended Data Figure 8: Pressure measurements from 04:00 – 12:00 UTC from Tonga, ~64km from Hunga Tonga. Note the multiple explosions after the initial primary Lamb wave trigger.



Extended Data Figure 9: Excess of H_2O (difference between the observation and the zonal mean) measured by IASI-B and IASI-C over Tonga at 30, 20, 10 and 2 hPa.