Theory and observation of acoustic coupling between the solid Earth and its atmosphere

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Abstract. Acoustic coupling between solid Earth and atmosphere has been observed since the 60^s on either ground-based or satellite measurements. This coupling leads to clearly identified signals on seismic data after atmospheric events or on atmospheric or ionospheric measurements after big quakes. Normal modes theory for the whole system Earth+atmosphere is well-adapted for the study of such signals. We present here the theory developped to compute realistic normal modes and the applications to modelling of coupling phenomena. Synthetics for either solid or atmospheric source or measurement are presented, showing a good agreement.

1 Introduction

Since about 10 years, many observations of coupling between the Solid Earth and its atmosphere were reported in seismology. At least three types of observations are supporting such coupling : The first one consists in the observation of seismic signals associated to strong atmospheric sources. Among the first were the records performed at Irkoutsk following the atmospheric explosion of a meteor or comet in Siberia on June, 30, 1908 [*Ben Menahem*, 1975]. More recently, seismic signals were recorded by *Cevolini* [1994], following an meteor atmospheric explosion in Italy. Similar signals were recorded following the major nuclear explosions performed by United States and Soviet Union in the atmosphere, between 1945 and their limited interdiction in 1963. Apart many reports and studies concerning atmospheric waves, [*e.g. Yamamoto*, 1956-1957, *Hunt*, 1960, *Press & Harkrider*, 1962, *Donn & Ewing*, 1962a-b, *Harkrider*, 1964] Other powerfull atmospheric sources are the volcano explosions, like El Chichon in 1982, and especially Pinatubo in 1991. By stacking 12 IDA stations during 12 hours, *Zurn & Widmer* [1996] have shown that the signals recorded following the Pinatubo eruption shown an selective excitation of Rayleigh surface waves around frequencies of 3.7 mHz, 4.44 mHz for the two mains peaks and 5.2 mHz, 6.1 mHz et 7.2 mHz. Many papers were published on the explanation of these unusual signals. Some have proposed a feedback regime between the atmosphere and the volcanoe [*Widmer & Zurn*, 1992, *Zurn & Widmer*, 1996]. Other proposed the excitation of two atmospheric waves, the low frequency one being a gravity wave, and the other being acoustic [*Kanamori & Mori*, 1992, *Kanamori et al.*, 1994].

The discovery of the continuous excitation of normal modes [*Suda et al.*,1998, *Kobayashi & Nishida*, 1998, *Tanimoto et al.*, 1998] has shown a second example of coupling between the solid Earth and its atmosphere. Such excitation appears to be produced by the turbulences of the Earth atmospheric boundary layer, which for wind extend within the first km of the surface and despictes seasonal variations [Nishida et al., 2000]. Simplified theory were proposed by *Tanimoto* [1999]. None are however able to fully accound for some features of the amplitude spectrum of normal modes.

The third example of coupling is related to ionospheric perturbations after earthquakes. Such a coupling can either be observed near a seismic source or at teleseismic distances. Following a Magnitude 5.9, *Kelley et al.* [1985], reported thermospheric perturbation of about 300 K between 300 and 400 km of altitude, *i.e.* corresponding to 25% of relative temperature variations. *Calais & Minster*, [1995], reported ionospheric perturbations by using GPS data. As shown by *Lognonné et al.*, [1998], a small fraction of the surface waves is transfered in the atmosphere for frequencies higher than 4.4 mHz. The surface waves therefore produce a plume, which can reach the high atmosphere after being amplified by the exponetial decay of the atmospheric density. *Artru et al.* [2001] have shown than such a signal might be therefore amplified by a factor of 50 000 when it reachs an altitude of 200 km. Typically, the detection threshold for atmospheric perturbations is about 10 m/s at 200 km (which generated for example Doppler effect of the order of $3x10^{-8}$). Such signal can therefore be detected for ground velocities of about 0.2 mm/s. For low

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magnitude quakes, such amplitudes are found only in the vicinity of the quake and the ionospheric perturbation is just above the epicenter. For large and very large earthquakes however, such amplitudes migh be found at teleseimic distances for the surface waves. Many observations were reported after large quakes in Alaska or Japan [*Yuen et al.*,1969, *Weaver et al.*, 1970, *Leonard & Barnes*, 1965, *Davis & Baker*, 1965]. New generation HF sounders are now able to monitor such signals for quakes larger than 6.5 in magnitude.
We recall in this paper the theory able to take into account these coupling effects by an explicit calculation of the normal modes of Earth model with realistic atmospheric model [*Lognonné et al.*, 1998]. We then show two applications for this theory: the first one is the analysis of the Pinatubo eruption, which can be foreseen as an example of excitation of the solid earth with source in the atmosphere. The second one is the recording of surface waves in the ionosphere, and therefore the excitation of atmosphere by earthquakes.

2 Normal mode theory: brief description

The atmosphere of the Earth (or other planets) changes the boundary conditions of the elastodynamic operator. The atmosphere is indeed such that no specific boundary can be defined, due to the exponential decay of the density. Moreover, acoustic waves are not reflected when propagating upward at high altitude, and loose their energy due to viscosity and non-linear effects.
In order to take into account these effects, it is first necessary to use a radiative boundary condition instead of the usual free surface boundary condition. Following *Unno* [1989] and *Watada* [1995], we assume a local dependence of the modes at the top of the atmosphere as r^{λ}. As shown by *Lognonné et al.,* [1998], each eigenfrequency ω determines two values for λ, respectively associated to modes with upward and downward propagating energy. This can be done with a variational method, which uses a basis of test functions, found by mapping the normal modes with free surface toward functions verifying explicitly the radiative boundary condition. It is also necessary to take, at higher altitude, above 100 km, viscous and possibly non-linear and thermal effects. The typical frequency domain of the ionospheric perturbations is from 1 to 50 mHz. In this range, viscous dissipation is expected to be important above 100 km high [*Pitteway & Hines,* 1963]. The viscous stress tensor can be expressed by:

$$\mathbf{T}'_{ij} = i\omega\mu_{vis}\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} - \frac{2}{3}\delta_{ij}\left(\nabla\cdot\mathbf{u}\right)\right). \quad (1)$$

Here, u_i is i^{th} component of displacement and μ the dynamic viscosity. This is a second order, frequency-dependent term that can be introduced in the variational process described in the previous section.

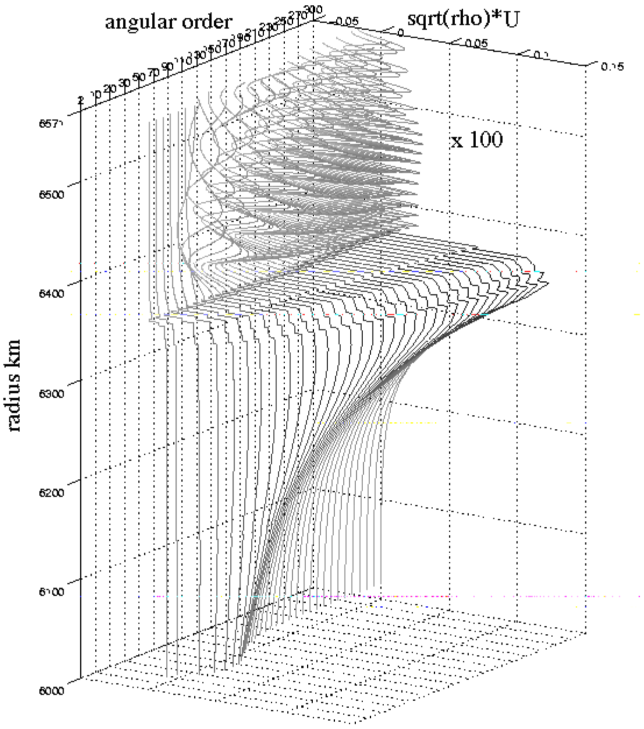

Fig. 1. Amplitude of the Solid spheroidal normal modes in the upper mantle and atmosphere.

We use(then the method of *Lognonné et al.*, [1998]. and found all normal modes of the Solid Earth-Ocean-Atmosphere coupled system (spheroidal solid Earth modes, acoustic, gravity and Lamb atmospheric modes, gravity-tsunami modes). The main perturbation, for the spheroidal normal modes are found in the amplitude of normal modes rather than in the frequency or quality coefficient, whose perturbations appears too small to be detected. Two regimes for the fundamental spheroidal modes are found: below 3.68 mHz (the exact frequency depending on the model), the atmospheric part of the mode is trapped and decreases exponentially with altitude. At higher frequencies in contrary, the energy propagates upward [*Fig.1*].

3 Seismic sources and seismograms

As shown in the previous section, Normal modes can be computed for a complete Earth model. The Normal mode summation techniques can then be also applied for the computation of seismograms, wherever is the source, *i.e.* in the atmosphere or in the Solid Earth.
For a source in the solid Earth, it can as usual be expressed with a Moment tensor. For a source in the atmosphere, *Lognonné et al.* [1994] have shown that the generalization of the concept of stress glut can be done. We recall here the demonstration. Let us consider the non-linear equation of momentum conservation, written

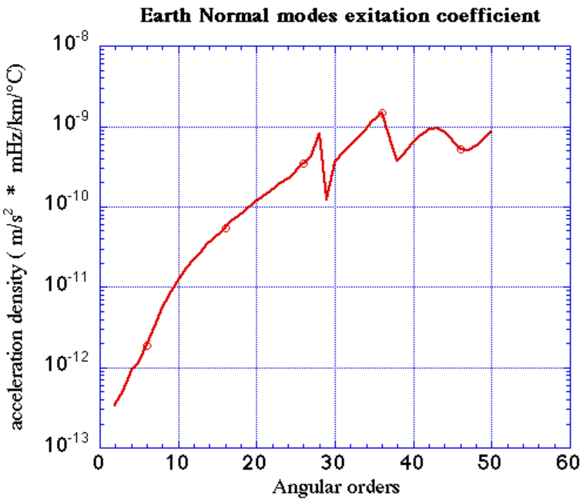

Fig. 2. Trace of the acceleration term for an isotropic atmospheric sources $A_{\ell,n} = \delta_{\alpha,\beta}\ A^{\alpha,\beta}_{\ell,n}$ at the Earth surface (z=0). Note the small amplifications near the two frequencies corresponding to the fundamental and first overtone of the atmospheric modes.

in the form

$$\partial_t(\rho\ \dot{\mathbf{u}}) = -\nabla p_{true} - \nabla.(\rho\dot{\mathbf{u}}\dot{\mathbf{u}}) \qquad (2)$$

where ρ is the density, $\mathbf{u}$ the displacement with respect to the equlibrium position, $\dot{\mathbf{u}} = \partial_t\mathbf{u}$ the velocity, p_{true} the pressure, and where we neglect the gravity terms for simplicity. This expression is the one valid in the atmosphere. On the other hand, let us take the equation governing linearized seismology, where for sake of simplicity, we neglect self-gravitation

$$\partial_t(\rho\ \dot{\mathbf{u}}) = -\nabla p_{hooke} + \mathbf{f} \qquad (3)$$

where the atmospheric force is $\mathbf{f}$, the model of Hooke pressure is defined as $p_{Hooke} = -\kappa\nabla.\mathbf{u}$ and where $\kappa = \gamma p$ is the bulk modulus of the fluid. The difference between the two equations allows us to express the excitation force, which is given by

$$\mathbf{f} = -\nabla.\mathbf{\Pi} \qquad (4)$$

where the momentum flux-glut tensor $\mathbf{\Pi}$ is the difference between the model momentum tensor, based upon Hooke's law and the true incremental momentum tensor. Here $\mathbf{\Pi}$ is given by :

$$\Pi_{ij} = (\mathrm{p}_{true} - \mathrm{p}_{hooke})\delta_{ij} + \rho\dot{u}_i\dot{u}_j. \qquad (5)$$

where δ_{ij} is the Kronecker symbol. The normal mode summation techniques [*Lognonné*, 1991] can now be used, below in a spherical model for sake of simplicity. Note that the theory used must take into account anelasticity, both related to the attenuation in the solid Earth and to the energy escape related to the radiative boundary condition. We finally obtain

$$\mathbf{u}(t,\mathbf{r}_s) = \sum_{k>0} \Re e\left(\frac{1}{i\sigma_k}\int_0^t dt' M_k(t')e^{i\sigma_k(t-t')}\mathbf{u}_k(\mathbf{r}_s)\right) .(6)$$

where $\mathbf{r}_s$ is the receiver/station location, index k denotes a given mode with quantum numbers ℓ,m,n , σ_k and $\mathbf{u}_k$ are the normal frequency and normal mode respectivelly associated to the index k and where the source term $M_k(t')$ is given by the source integrated over the whole source volume and is expressed by

$$\begin{aligned} M_k(t) &= \int dV\ \Pi_{ij}\nabla^i v_k^j \\ &= \sum_{\alpha,\beta}\int dr\ r^2 E^{\alpha\beta}_{\ell n}\int d\Sigma\ Y_\ell^{Nm}\ \Pi_{\alpha\beta}\ , \end{aligned} \qquad (7)$$

where we use a decomposition of the source term in generalized spherical harmonics and where $\mathbf{v}_k$ is the dual normal mode. For global scale atmospheric sources, as those related to the continuous excitation of normal modes, all terms of (5) are important, in contrary to the hypothesis of *Tanimoto* [1999] which consists in neglecting the first part. Moreover, most of the pressure variations in the atmosphere are non-acoustic, i.e. such that $p_{hooke} << p_{true}$. Typical pressure spectral amplitude in the frequency range of normal modes are of the order of 0.5 $mbar/\sqrt{Hz}$ [*Beauduin et al.*, 1996], equivalent to a square velocity spectral density of $6.5^2\ m^2/s^2/\sqrt{Hz}$. For sources in the water, the pressure glut must be computed by using the difference between the two pressure terms. In both cases, the ground acceleration can be expressed as

$$\begin{aligned} a(t,\mathbf{r}_s) &= \sum_{k,\alpha,\beta} Y_\ell^{0m}(\theta_s,\phi_s)\ \Re e\int_0^t dt'\int dz \\ & A_k^{\alpha\beta}(z,z_s)\frac{M^k_{\alpha\beta}(t',z)}{\rho(z)R}ie^{i\sigma_k(t-t')} \end{aligned} \qquad (8)$$

where we choose to represent the source terms in temperature units integrated for altitudes $z = r - r_s$, and where

$$\begin{aligned} A_k^{\alpha\beta}(z,z_s) &= R\sigma_k u_k(r_s)\rho r^2 E^{\alpha\beta}_{\ell n}(r), \\ M^k_{\alpha\beta} &= \int d\Sigma\ Y_\ell^{Nm}\ \Pi_{\alpha\beta}. \end{aligned}$$

Here R is heat constant by mass unit and u_k the vertical amplitude of the mode. The trace of the acceleration density is given in *Fig. 2.*

4 Pinatubo eruption

As a first application we consider the atmospheric source term of the Pinatubo eruption. We take 18 stations of the Global Network (Geoscope and Iris) on the VLP channels corresponding to the full day of June, 15, 1991.

Two small quakes recorded on the data and originating from other sources as the Pinatubo region are substracted from the data after a waveform fitting. The data cleaned are shown on the top of *Fig. 3.* We then perform a least square inversion of these data with synthetics filtered in the frequency bandpass window from

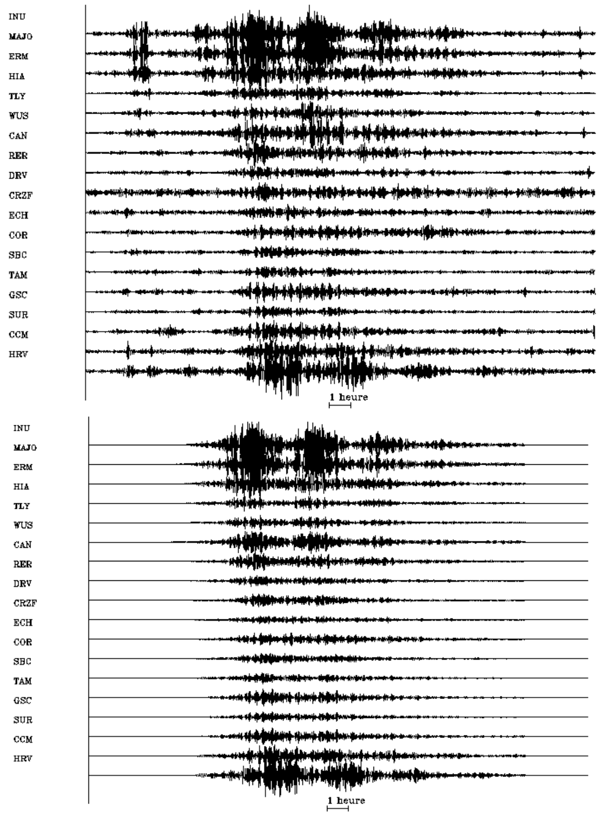

Fig. 3. Top, bandpass filtered data recorded after the Pinatubo eruption by several station of the global network. Bottom: synthetics found in the study

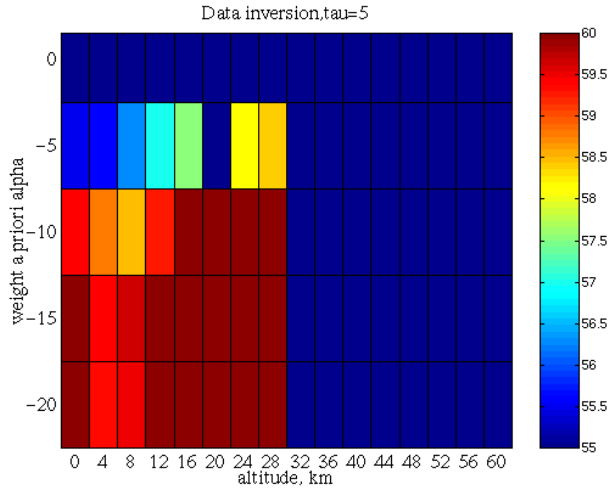

Fig. 4. variance reduction for a serie of inversion, for different values of altitude and weighting factor ϵ.

1 mHz to 8 mHz, assuming that the seismic source is localized at a given altitude / depth z, is isotropic in direction and can radiate during 10 hours starting after June, 14, 1991. From relation (8), we then compute the acceleration at the stations altitude z_s as

$$a(t,\mathbf{r}_s) = \sum_k \Re e \frac{Y_\ell^{0m} A_k(z,z_s)}{\rho(z)R} \int_0^t dt' \quad m_k(t',z)\, e^{i\sigma_k(t-t')}\ , \qquad (9)$$

extending the summation up to the 10^{th} harmonics and for all angular orders in the frequency window.

The inversion is performed by least square fitting of the vertical ground displacement after instruments correction and by adding a correlation time to the moment tensor history, and therefore by minimizing

$$Cost = \sum_n \int (u^n_{obs}(t) - u^n_{cal}(t))^2 + \epsilon \int dt\ m(t) C^{-1}(t-t') m(t'). \qquad (10)$$

We chose an exponential correlation function $C(t) = e^{-\frac{t^2}{\tau^2}}$ to stabilize inverse problem and performed inversions for all altitudes from a few kilometers depth to about 60 km of altitude. The best variance reduction (about 60 %, see *Fig. 4*) is found near the surface and at an altitude between 24 and 28 km.

In order to assess the validity of a low altitude source with respect to a high altitude one, we compared the amplitudes of the two different sources. The seismic moment of the source ranges between a minimum expressed as $M_0 = \tau_b(\gamma - 1)E$ [*Lognonné et al.*, 1994], where γ is the adiabatic index of the atmosphere and τ_b the duration of the blast, and $M_0 = 2\tau_b E$ when all the energy is released in keinetic energy, which might be the case for the eruption where most of the ejecta have a vertical velocity. As shown by *Fig. 5*, reasonable amplitudes are found only for a source at 24-28 km of altitude, with most of the energy released at the time of the individual explosions and release of seismic moment are found near the reported date of the individual eruptions. These eruptions are associated to yields of about 4000 MT.sec, corresponding to explosion releasing about 20 MT during blast times of about 200-500 sec, which corresponds to the order of magnitude of the Pinatubo eruption, which released about 200 MT of energy in several explosions. Our results show that the seismic source of the Pinatubo eruption can be relatively well explained by a series of eruption rather than the complex mechanisms proposed by the previous studies.

5 Atmospheric signals from seismic sources

As a second application, we show atmospheric signals produced by an earthquake. We present for that purpose data of ionospheric oscillations recorded in France by an ionospheric sounder developed by *Commissariat à l'Énergie Atomique*. This instrument performs the measurement of the Doppler shift between a HF EM wave emitted from the ground and its counterpart reflected by ionospheric F layer, and provides a direct measurement

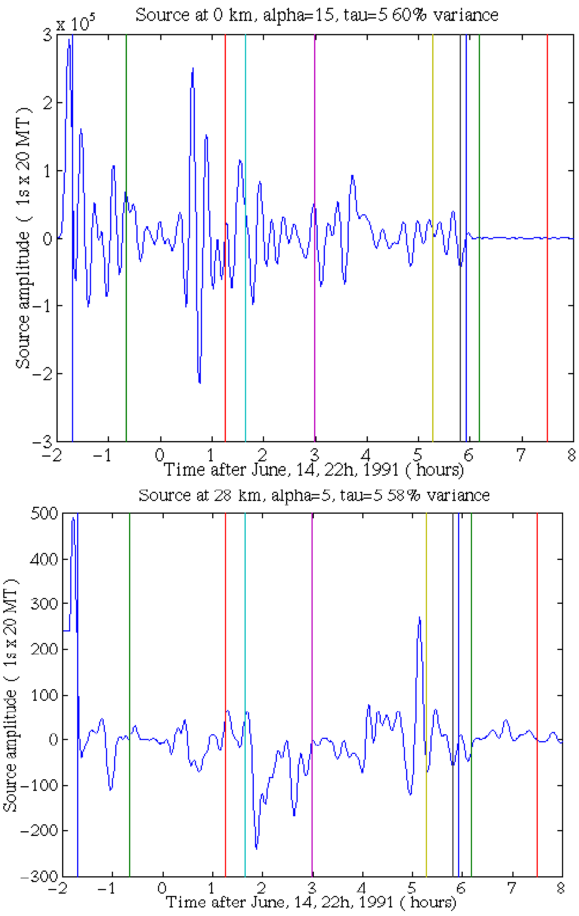

Fig. 5. Top, source history for a surface pressure glut versus time. Amplitude is in 20 MT of equivalent TNT times one second. Bottom: same, but for a source at 28 km of altitude. Note that both the amplitude of the source and the complexity is reduced. The source term is closer from a series of explosion, each of them of about 20-40 MT and with burst times of the order of 200-500 sec. Vertical lines are associated to the reported eruption of the volcano.

of the vertical velocity $\mathbf{v}$ of the ionosphere, following the equation:

$$\delta f = -2 f_0 \frac{\mathbf{v}\ \cos(\theta)}{\mathbf{c}}, \tag{11}$$

where θ is the zenithal angle of the ray.

The reflection altitude depends on the electron density profile : reflection occurs when plasma frequency is equal to wave frequency. The CEA network consists in 3 receivers,located 50-100 km apart from the emitter in Francourville, France. The network has been working continuously since August 1999, most of $M > 6.5$ earthquakes have been observed [*Table 1*]. We present the comparison of synthetics with these data [*Figure 6*]. Significant differences in amplitude are found. Even if non-linear effects are not taken into account here, they should represent about 10% of the signal at 150km and can be neglected in first approximation. In contrary, we found that attenuation due to viscosity is important

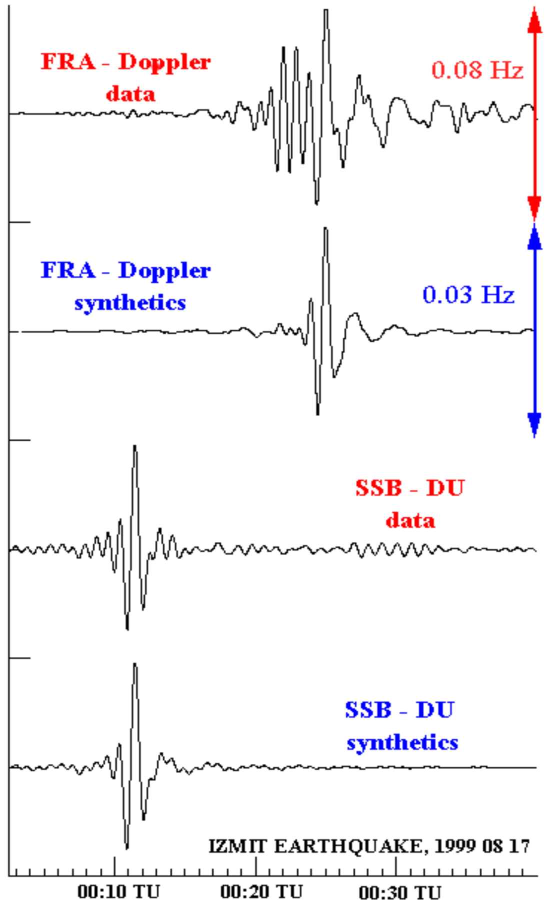

Fig. 6. Observation of ionograms at Francourville and seismograms at the nearby Geoscope SSB station, with their synthetics for the PREM model with US standart atmospheric model.

above 100 km height and strongly constraints the attenuation of waves: our data can therefore be used to constraint the relatively poorly known profiles of the atmospheric viscosity. Future works will also combine this method with the modelling of the interaction of the wave with ionospheric plasma, together with thermal effects.

Date	Earthquake	Ms	Depth (km)	Δf (Hz)	Alt (km)	T (s)
8/17/99	Turquey	7.8	10	0.5	240	100
8/20/99	Costa Rica	6.7	33	0.4	170	40
9/20/99	Taiwan	7.6	10	0.3	186	70
9/30/99	Mexico	7.5	33	0.15	170?	100
10/16/99	California	7.3	10	0.7	182	40
11/12/99	Turquie	7.2	10	0.4	221	40
3/28/00	Japan	7.7	116	0.8	162	50
5/4/00	Minahassa	7.3	33	0.3	280	100
6/4/00	Indonesia	7.9	33	0.4	168	50
6/18/00	Indian Ocean	7.5	10	1.05	169	

Table 1. List of the events recorded during the first year of operation. Amplitude of the doppler signals are given, as well as the altitude A and main period T of the measurement, for all recorded events.

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