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Coupled modes at interfaces: A review

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RESUMEN

El acoplamiento entre ondas puede producirse cuando un pulso acústico selecciona un modo de Rayleigh de la misma velocidad, y ambos se propagan juntos intercambiando energía a través de la interface. Se observó una señal acoplada en sismos provenientes de las fracturas de Blanco y Mendocino, frente a las costas de Norteamérica, registrados en el Observatorio Hawaii-2 en el fondo oceánico. La señal, con trayectoria puramente oceánica, parece consistir en modos superiores de Rayleigh superpuestos y no dispersados, que se propagan sobre el fondo marino tanto en el sedimento como en el agua. Estos modos acoplados se distinguen por su composición en frecuencia, y por sus velocidades de fase y de grupo. El acoplamiento sismoacústico se produce bajo las siguientes condiciones: a. hay una interface de baja velocidad en el piso submarino, b. La longitud de onda de las componentes de Rayleigh es menor que la profundidad del agua, y c. Existe una no linealidad débil en la relación esfuerzo-deformación en la interface.

PALABRAS CLAVE: Acoplamiento de ondas, ondas sísmicas no lineales.

ABSTRACT

Wave-to-wave coupling may arise when an acoustic pulse selects a Rayleigh mode of the same speed and both travel together swapping energy across an interface. A distinctive coupled signal called *Ti* was observed at the Hawaii-2 Observatory from earthquakes on the Blanco and Mendocino Fracture Zones, off the coast of North America. The signal travels along a purely oceanic path; it appears to be a composite of undispersed higher Rayleigh modes propagating along the ocean floor both in the sediments and in the water. Coupled modes may be identified by their frequency composition and by their phase and group velocities. Seismoacoustic coupling at the seafloor is conditioned on (a) the presence of a low-velocity interface at the ocean floor, (b) the wavelength of the Rayleigh mode being shorter than the depth of the water layer, and (c) weak stress-strain nonlinearity at the interface. It is conjectured that coupled interface waves may exist at other interfaces, including the Moho and the core-mantle boundary.

KEY WORDS: Ware-wave coupling, nonlinear seismic waves.

INTRODUCTION

The Plains Indians of North America detected the approach of enemy horsemen by holding an ear close to the ground. The acoustical signal from the percussion of the hoofs was audible near the air-soil interface. This effect may be representative of a variety of coupled interface signals that depend on weak nonlinear interactions between modes.

An early example was a Rayleigh-acoustic mode known to exploration geophysicists as *ground roll* (Press and Ewing, 1951; Ewing, Jardetzky and Press, 1957, referred to below as "EJP"). This monochromatic wave train was described as an "air-coupled Rayleigh wave". Here we comment on recent observations of similar seismoacoustic waves found on the Pacific ocean floor (Butler and Lomnitz, 2002). These coupled interface modes are typically undispersed and of finite duration. Because of coherence they can propagate efficiently with little attenuation, like laser beams. Physically they may be seen as hybrids between higher Rayleigh modes traveling at stationary values of the group velocity, and body waves, but they are peculiar in that the phase velocity in one medium is pegged to the speed of the acoustic wave in the other medium. In this contribution we review what is known about the physics of coupled modes, and we advance some theoretical considerations and some conjectures on the role of wave-to-wave coupling in seismology.

SEISMOACOUSTIC MODES ON THE OCEAN FLOOR

A distinctive coupled signal called *Ti* was identified on seismic and hydrophone records at the Hawaii-2 Observatory (H2O), located on the Pacific ocean floor east of Hawaii at a depth of 4979 m (Butler and Lomnitz, 2002, referred to below as "BL"; Butler *et al.,* 2000). The Hawaii-2 observatory is a complete, self-contained seismological laboratory recording in real time on the deep ocean floor. The signals are transmitted to Honolulu by submarine cable. Records at H2O of purely oceanic events present a unique character.

Figure 1 shows a seismogram of the earthquake of 2 June 2000 (M=5.9) on the Blanco Fracture Zone, off the coast of northern California. Similar records have been obtained, mostly from events on the Blanco and Mendocino Fracture Zones but also from the coast of North and Central America. The path is fully oceanic and the arrivals are pulse-like. The unusual signal at 1430 s after origin time corresponds to a group velocity of 1510 m/s, the speed of sound in deep water.

A similar signal has been routinely identified as a *T* wave, an acoustical signal that propagates in the SOFAR channel located between 100 m and about 2.5 km depth (Okal, 2001). Both signals propagate at the speed of sound in water; however, there is no direct connection between station H2O and the SOFAR channel. Also, *T* waves should be distinctly slower than 1510 m/s because the SOFAR channel is a low-velocity waveguide in the ocean. Finally, the *Ti* signal was recorded on seismic sensors buried at 0.5 m below the ocean floor as well as on the hydrophone tethered at 0.5 m above the ocean floor. Both records contain discrete frequencies of the acoustic signal, corresponding to higher Rayleigh modes (Figure 2, reproduced from BL).

In Reference BL, this signal was identified as a coupled seismoacoustic wave, by eliminating other possibilities. Thus, a *T* phase is a purely acoustic guided signal in the upper ocean: even if it could propagate on the ocean floor it cannot select frequencies corresponding to specific Rayleigh modes. On the other hand, a mere superposition of Rayleigh

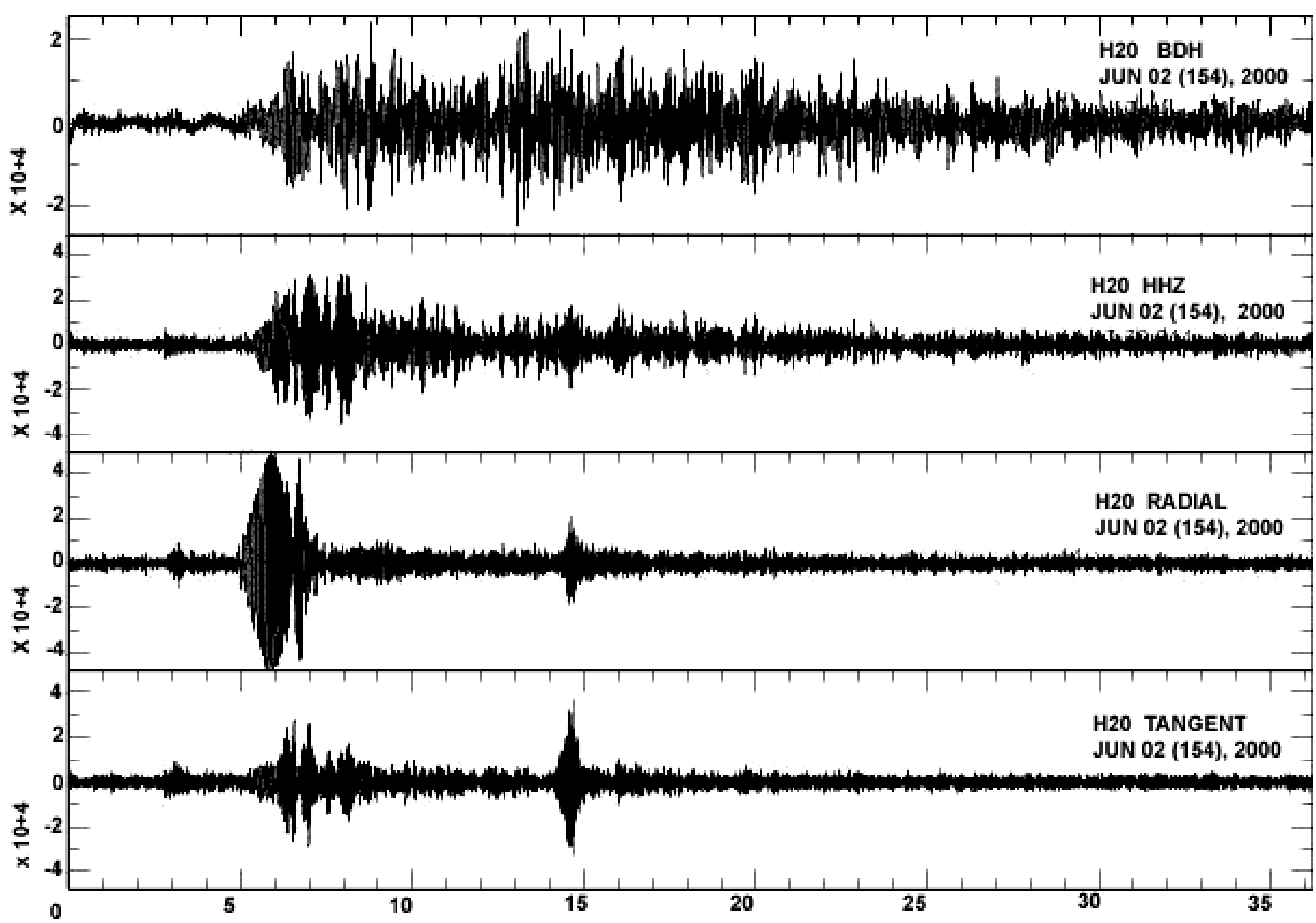


Fig. 1. Seismogram of the earthquake of June 2, 2000 (M6.2) on the Blanco Fracture Zone recorded at ocean-bottom station H2O (27.88ºN, 141.99ºW). The epicentral distance is about 2160 km. The upper trace shows the hydrophone record; next comes the vertical component of velocity at a seismometer buried 0.5 m below the sea floor at 4979 m depth. All amplitudes plotted are 6 dB above the background noise prior to *P* arrival. Time is in hundreds of seconds after origin time.

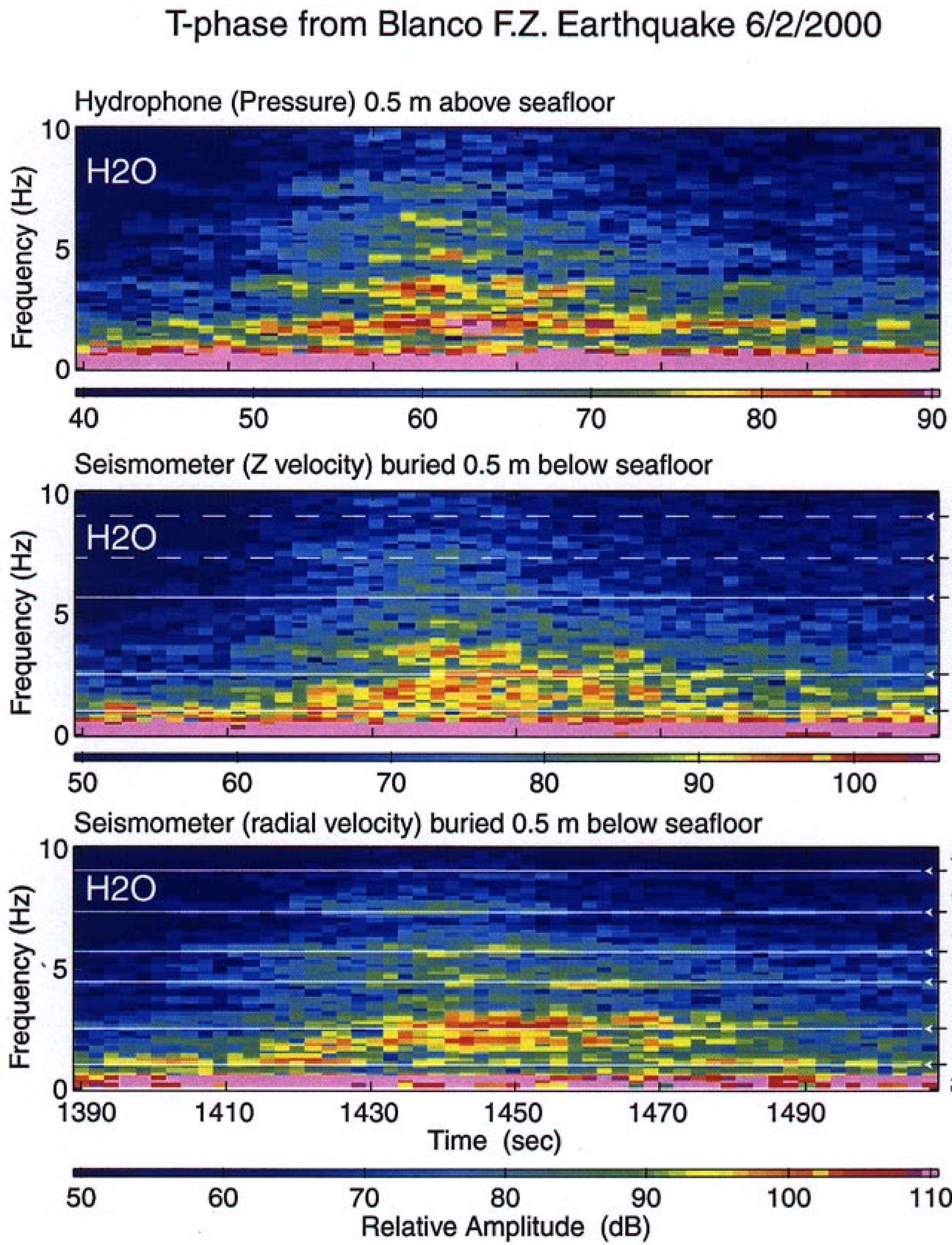


Fig. 2. Spectrogram of the hydrophone (*bottom*) and seismic signals (vertical components, top) arriving about 1400 s after origin time of the earthquake in Figure 1. Notice the banded modal structure and the absence of dispersion.

modes should show dispersion. The signal arrives at a phase velocity corresponding to the speed of sound in water: the acoustical signal in the water is coupled to the elastic wave in the bottom. A similar conclusion was reached by EJP with regard to air-coupled Rayleigh waves.

Other possible explanations of the observed signal included shear modes in sediments (Coulomb and Molard, 1949; Leet *et al.,* 1951), Scholte waves (Nolet and Dorman, 1996), Stoneley modes (Biot, 1952), Rayleigh modes (Okal and Talandier, 1981; Sykes and Oliver, 1064), and mode scattering (Park *et al.,* 2001). None of these would generate a signal that matched the distinctive features of the spectrogram of Figure 2.

The frequency bands on the hydrophone record and on the seismic record match exactly. While *T* and *Ti* phases travel at similar speeds their modal composition is different. Converted seismic signals identified as "*T phases*" at land-based seismic stations, especially on Pacific islands, might be caused by *T* or *Ti* or both.

WAVE-TO-WAVE COUPLING

Ewing, Jardetzky and Press (1957) first described wave-to-wave coupling between a pulse in a fluid medium and a Rayleigh wave in the ground. They introduced the idea of coupling by invoking Huyghens' Principle:

"The traveling impulse may be replaced by a succession of infinitesimal impulses placed at equal intervals of time along the path of the disturbance. Each impulse initiates a train of dispersive waves, and constructive interference is possible only for those waves whose phase velocity c equals the speed of the traveling disturbance c_0. The energy thus transferred will form a train of constant-frequency waves. The duration of the wave train at any distance will be proportional to $|1/c_0 - 1/U_0|$, where U_0 is the group velocity corresponding to c_0" (EJP, p. 230-231).

As Lamb (1916) may have realized, this mechanism implies the existence of weak nonlinearity. In a linear non-dissipative system different modes co-exist without interacting: they are superposed. Thus it is necessary to introduce a weak dissipation, presumably due to the roughness of the interface. The duration of the wave train was correctly given by EJP but was incorrectly attributed to a nearly undispersed new "branch" of the group-velocity dispersion diagram. We have been unable to confirm the presence of such a new branch and it is unclear how it might arise.

Here is what one finds instead. In linear Rayleigh theory the dispersion relations for a layered halfspace are obtained from the roots of a secular function F computed from the determinant of order n of the boundary conditions:

$$F(k,c) = P^{2(n+1)}\Phi_{n+1} + P^{2n}\Phi_{2n} + \ldots + P^{0}\Phi_{0}\,, \tag{1}$$

where k is the wavenumber, c is the phase velocity, and n is the number of layers. The P^{2i} are even polynomials of degree $2i$ in c and the Φ_i's are hyperbolic functions in k. The roots $c=c_m(k)$, $m=0, 1, \ldots$ of Eq (1) are the dispersion relations for the modes m.

Thus the function F is strongly nonlinear and unsurprisingly its properties have not been fully explored. The top layer—water in the case of the ocean floor—was air in the case discussed by EJP. When the fluid is a halfspace it turns out that the secular function F becomes complex-valued for all $c>c_0$, where c_0 is the speed of sound in the fluid layer. At exactly $c=c_0$ the secular function has a singularity.

Thus the group velocity U must develop a gap in the interval between $U=c_0$ and some finite value $U=U_0$—precisely the group velocity corresponding to c_0. This gap is bridged by the coupled interface wave train. This is best shown by a numerical example. Figure 3 is a bitmap of Re$|F(k,c)|$ computed for station H2O for (a) a finite water layer (h=4.979 km, *top*), and (b) a water halfspace ($h=\infty$, *bottom*). A simple 3-layer model of the ocean floor was used (Table 1). For a water halfspace the phase space becomes complex at c>1510 m/s, the speed of sound in water (*bottom figure*). The first real phase velocity occurs precisely at c_0=1510 m/s. As the group velocity U_0=1455 m/s corresponding to this value is lower than c_0 there is a gap in the group velocities between 1455 m/s and 1510 m/s. When the water layer is finite (h=4.979 km), no such discontinuity develops.

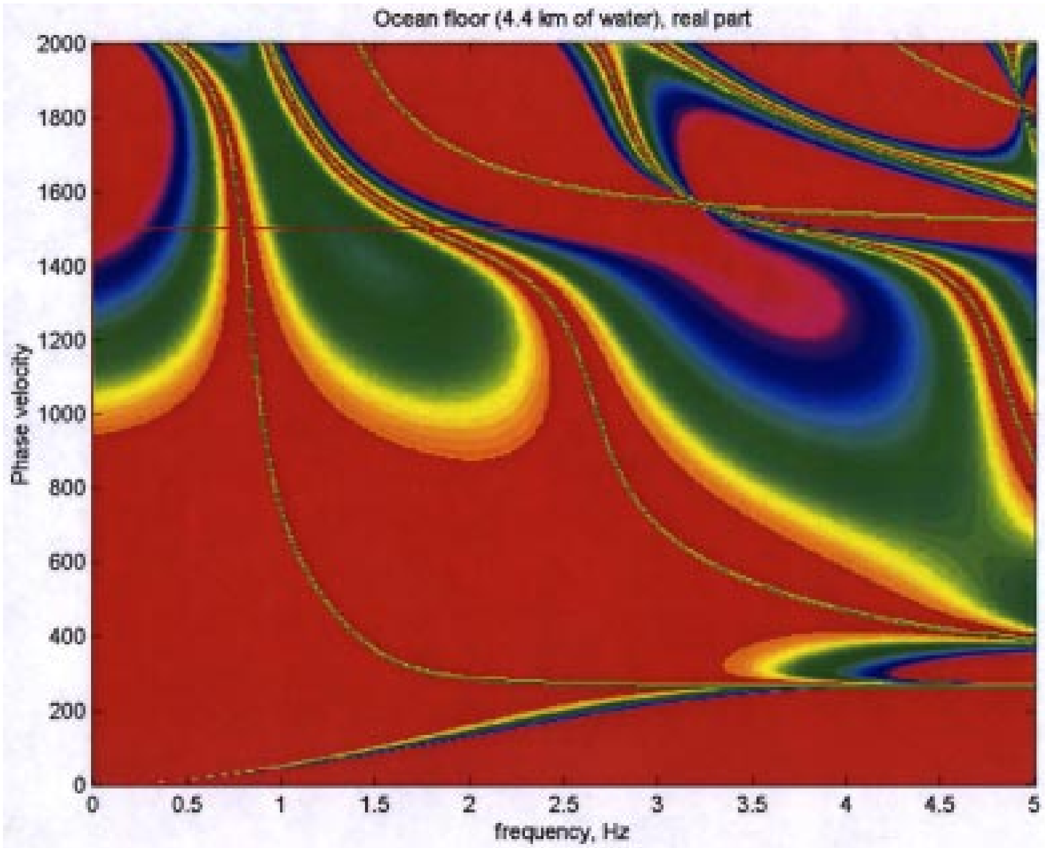
In Figure 4 the group velocities are shown as green lines. Note that the fundamental mode shows no gap—at least not on the scale of the figure. The wavelength of the fundamental mode at $c = c_0$ is on the order of 5 km which is about the depth of the water layer. Since the wavelength of the overtones is much shorter than 5 km these modes can-

Table 1

Three-layer ocean-floor model, broadly after Nolet and Dorman (1996)

Layer	Thickness, m	Vp, m/s	Vs, m/s	ρ, kg/m^3
Water	4979*	1510	0	1000
Mud	85	1550	329.67	1100
Rock	∞	4400	2200	1920

* For the deep-water case (Figure 3, *bottom*) the water layer is taken to be a half-space.



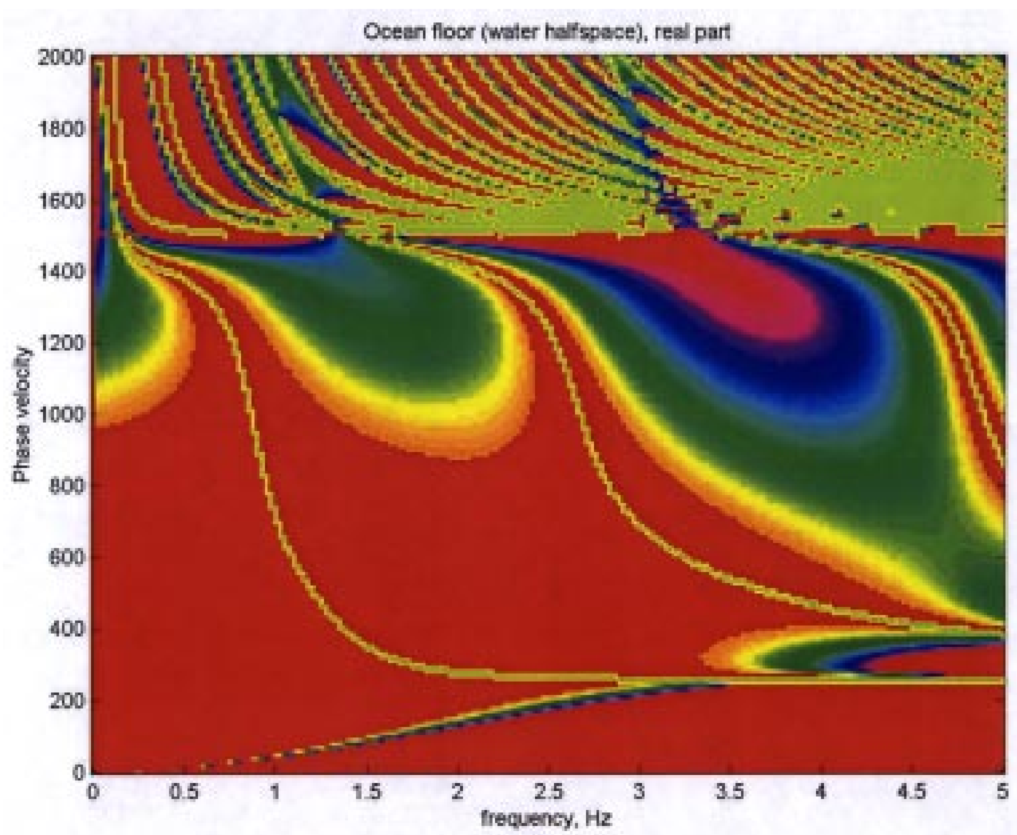


Fig. 3. A bitmap of the dispersion function $|F(k,c)|$ for a 3-layer model of the ocean floor (see Table 1). *Top,* shallow water, *bottom,* deep water (real part). *Yellow lines,* zero crossings of $F(k,c)$.

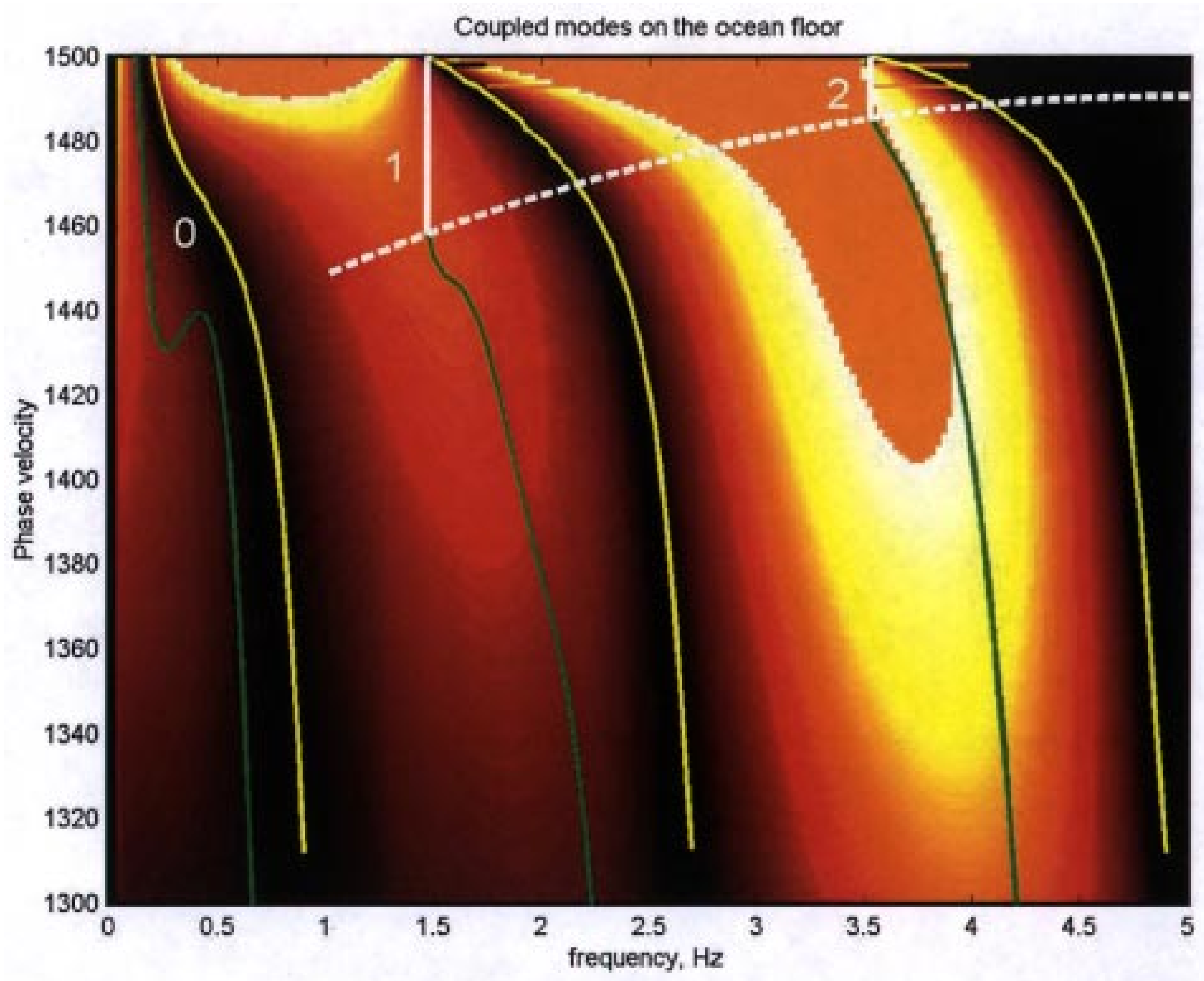


Fig. 4. Coupled *Ti* modes (*white lines*) for the first two overtones at the ocean floor. Parameters as in Figure 3 *(bottom*). *Yellow,* phase velocity; *green,* group velocity. The duration of the coupled signal (*dashed line*) decreases toward the higher modes.

not tell the difference between a finite layer and a halfspace of water. Hence coupling occurs only for the higher modes.

The frequencies of the coupled modes (*white lines*) are accurately predicted from the dispersion diagram of Figure 4. Note that these frequencies are 1.45 Hz (mode 1) and 3.53 Hz (mode 2), in excellent agreement with the frequency bands observed in the spectrograms of Figure 2. Note also that the duration of the signal decreases toward the higher modes (*white dashed line*, Figure 4). Again, this feature is observed in the spectrogram.

The white lines in Figure 4 were interpreted by EJP (p. 236), as "an additional train introduced by coupling . . . This train begins at a time corresponding to propagation at the speed of sound in [the fluid] and continues with almost constant frequency until the time $t=r/0.44\beta_1$," where β_1 is the shear-wave velocity in the solid layer. This interpretation agrees with ours. The point is, however, that each wavelet in this wave train propagates at a stationary value of the group velocity. It is the equivalent of an Airy phase (Pekeris, 1948). Each wavelet has a different group velocity. The hybrid nature of this wave train is manifested by the fact that its phase velocity remains constant (like an acoustic wave) while its group velocity decreases in time (like a Rayleigh wave).

The first such interface mode—a special case of generalized Rayleigh wave—was discovered by Stoneley (1925). But there remains a whole family of coupled interface modes to be discovered.

INTERACTION BETWEEN EARTH MODELS

Coupled interface modes may not be supported by all interfaces. In the seismo-acoustic case, a common mode must be shared by the acoustic wave and the Rayleigh wave. This condition implies that c_0, the speed of sound in the fluid layer, must fall within the range of available Rayleigh phase velocities. If β_{min} is the minimum shear-wave velocity in the stratigraphic column, we must have

$$0.91\beta_{min} < c_0 \quad (2)$$

for coupling to occur.

As we have seen, there is another condition on the wavelengths. In the present example the fundamental mode is not coupled because its wavelength exceeds the depth of the water and the condition $h=\infty$ does not apply.

But how can we explain the fact that both models appear to co-exist in the same seismogram? Indeed, the fundamental mode exists on the seismogram (but not on the hydrophone record!), over the entire domain of the secular function and not just for c<1510 m/s. Note, e.g., the prominent train of fundamental-mode Rayleigh waves arriving about t=500 s, with phase velocities of around 4 km/s (Figure 1).

Thus both structural models (for finite and infinite water layers) generate signals on the seismogram, and energy appears to be partitioned between these models. Coupling can only occur when the water layer is a halfspace, and the fundamental mode can only propagate when the water layer is finite. How can this duality be understood?

The answer must be sought in the co-existence of different structural models when the Earth is a complex nonlinear system. In this case, any structural model is an idealization with a different probability attached to it. We suggest that bitmaps of the secular function F may be thought of as probability maps for the existence of specific modes. If so, the yellow lines in Figure 3 (a and b) merely represent potential modes at some peak level of probability. A realization of a given mode cannot know whether the water layer is actually finite or infinite.

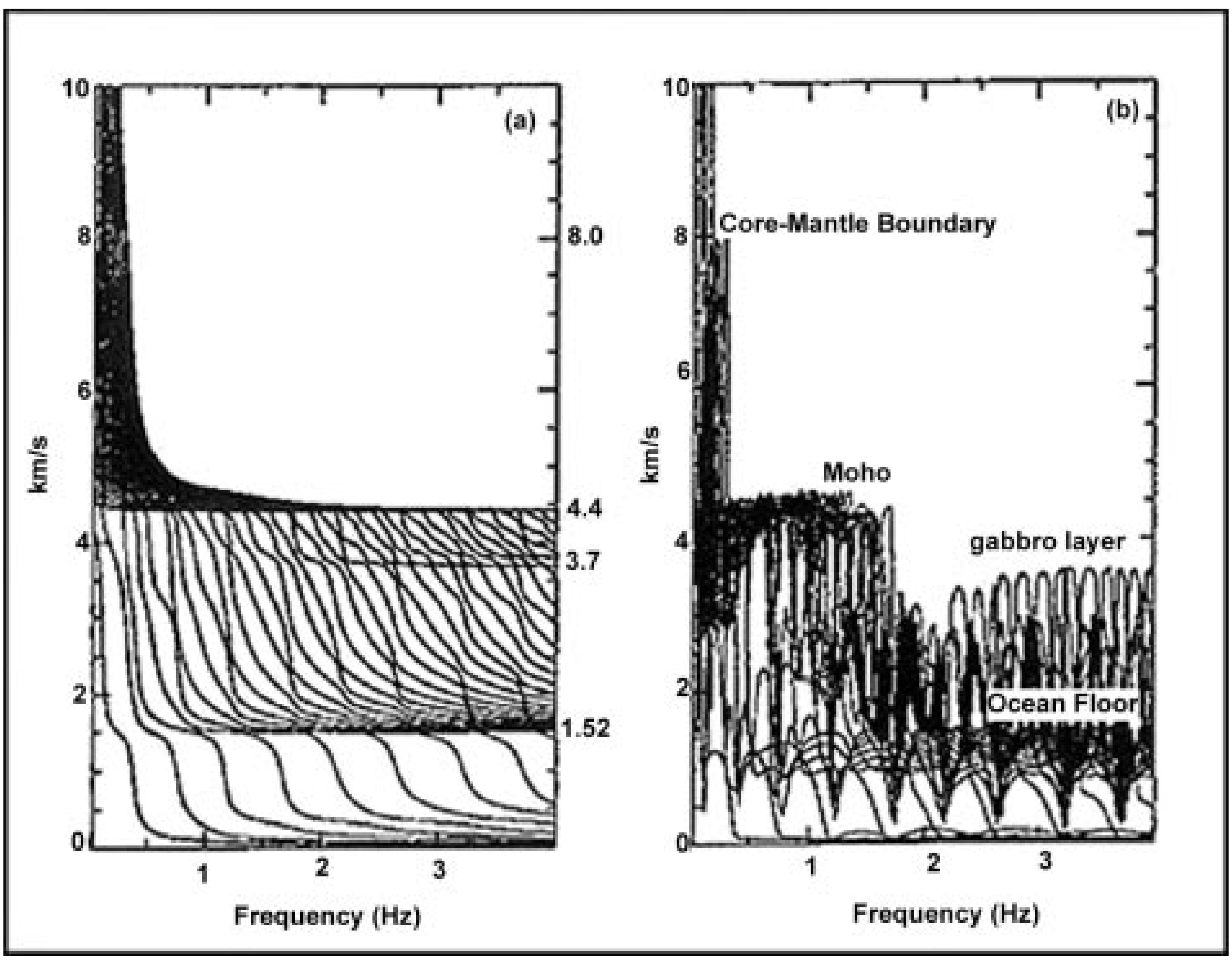


Fig. 5. Rayleigh dispersion curves for a standard Earth model with an oceanic layer, after Nolet and Dorman (1996). *Left,* phase velocities; *right,* group velocities.

If all modes are possible and if a wavelength is significantly shorter than the water depth the corresponding mode behaves as if the water layer was a halfspace. Otherwise the mode can sense the surface and it behaves as if the water layer was finite.

But the same modal uncertainty must also exist over a continental path. Suppose that a Rayleigh wave propagates from an epicenter near Acapulco to a station in Mexico City. The structural model varies continuously along the path, as each point is underlain by a different structure. The modal composition of the signal must change along the path. By the time the Rayleigh wave arrives in Mexico City, the modal content should correspond to some average structure. Instead, all the modes generated along the path are present in the signal.

EXISTENCE OF COUPLED MODES AT INTERFACES: A CONJECTURE

Coupling cannot be restricted to solid-fluid interfaces. It should occur at the Moho. Figure 5 shows the dispersion relations for a standard Earth model with an oceanic water layer (Nolet and Dorman, 1996). Notice the step-like structure of the modes. The phase-velocity dispersion curve for each mode contains $2n$ inflection points, corresponding to n maxima and n minima of the group velocity. Positive inflections (group-velocity maxima) correspond to major interfaces where coupling may occur.

Figure 5 predicts that the Moho should admit coupling to a body wave with a phase velocity of 4.4 km/s—the S-wave velocity in the upper mantle. We should search for a coherent multimodal signal with a finite time duration and a phase velocity of 4.4 km/s. Such a signal exists: it is the S-wave coda. Jeffreys (1970) was unable to explain the S-wave coda by appealing to dispersion, attenuation or scattering. Other explanations require the existence of large scatterers in the middle crust which have yet to be actually detected.

The S-wave coda detected at the earth's surface, of course, is not the coupled mode itself but the head wave it generates in the crust.

Back to Figure 5, other coupled modes might exist at the core-mantle boundary (c=8 km/s), and less prominently at the Conrad Discontinuity (c=3.7 km/s). As the Earth model is further refined we are likely to find other interfaces capable of admitting coupled modes, e.g. at the 600 km discontinuity and in the D” layer. Such modes may have actually been observed, e.g., *Sa* waves (Schwab *et al.,* 1974).

Pekeris (1948) showed that Airy phases propagating at an interface over an epicentral distance r present an amplitude gain by a factor of $r^{1/6}$ over neighboring Rayleigh modes. Thus an Airy phase will tend to emerge from the signal as it propagates. This is also true for coupled interface modes, except that an Airy phase is a single wavelet and a coupled mode is a succession of Airy phases in time.

NONLINEARITY

The role of nonlinearity in surface wave theory deserves some comment. As suggested by Infeld and Rowlands (1990), body waves in an infinite homogeneous space can be strictly linear, but the presence of an interface means nonlinearity. In particular, this applies to wave-to-wave coupling.

In the case of water waves, weak nonlinearity is introduced because of the fact that the displacements at the free surface cannot be neglected in the equations of motion. Elsewhere the nonlinearity may be present in the constitutive equations, as in the case of soils. Infeld and Rowlands (1990) show that the nonlinear Schrödinger equation yields good results for discussing the stability of wave trains and solitons, with applications in most fields of classical physics.

Coupled seismoacoustic signals on the ocean floor are easily converted to P waves when they encounter an island or a continental shelf. This explains the routine observations of converted *Ti* phases at seismic stations on land. According to Eq (2), however, *Ti* modes can only propagate when the seafloor lining has a shear-wave velocity below 1.5 km/s along the path. Coupled modes might be used as a diagnostic tool for the presence of mud or other low-velocity materials on the ocean floor. *Ti* has now been observed for an important number of earthquakes where the propagation path crosses the East Pacific Rise (see BL). There is little or no sediment cover in the axial valley of the ridge crest, but a sediment cover does exist where the ridge axis is offset by transform faults. Seismic studies of layer 2A at the top of this young crust indicate very low shear velocities (0.4-0.8 km/sec) attributed to high-porosity pillows and flows (Christeson *et al.,* 1994; 1997). Such materials would ensure continuity of propagation of *Ti* for paths across oceanic spreading zones. The structure of the ocean floor beneath the mud layer does not play a significant role in the propagation of *Ti*. The frequency content of the mode is determined by the average thickness of the mud layer. Observations of *Ti* are also seen for distant events (>9000 km) with dominantly pure-oceanic paths.

The influence of nonlinearity is also observed in the particle motion of the *Ti* signal. It is 3-D elliptical, with a significant transverse component (Figure 1). This cannot be due to Love waves—the frequencies would show up on the spectrogram of the transverse component. Elliptical 3-D particle motion is observed in other nonlinear wave, e.g. in water waves or whenever coherent monochromatic wavetrains propagate on very soft ground (Lomnitz *et al.,* 1999).

CONCLUSIONS

Exotic waves caused by coupling at interfaces may be more common than has been generally realized. In the case of the ocean-floor wave *Ti,* these coupled modes may have been mistaken for *T* phases. The mechanism of coupling appears to require weak nonlinearity, in addition to strict numerical conditions on the phase velocities and wavelengths.

Coupled interface waves can propagate efficiently over distances of thousands of kilometers without significant attenuation. At an epicentral distance of 2000 km the *Ti* phase emerges as the dominant phase in the seismogram above 1 Hz. Thus seismoacoustic *Ti* phases may become relevant in the context of the International Monitoring System (IMS) of the Comprehensive Test Ban Treaty Organization, which routinely uses observations of converted oceanic phases at land-based stations for monitoring oceanic seismicity.

T phases are routinely used for monitoring oceanic seismicity via SOFAR hydrophones (Fox *et al.,* 2001), and are now being utilized by the International Monitoring System (IMS) of the Comprehensive Test Ban Treaty Organisation. The IMS has SOFAR-channel hydrophones as well as seismometers located on islands. The observations at H2O suggest that *Ti* may be the most energetic seismic wave propagating at the sea floor at frequencies above 1 Hz. The conversion of *Ti* to *P* waves at the continental slope or an island may be an important component of apparent observations of converted *T* phases on land. Array measurements throughout the water column and within the sea floor would advance our understanding of the propagation of the propagation and conversion of these coupled seismoacoustic interface waves.

It is conjectured that coupled modes may also occur at other interfaces such as the Moho, the core-mantle boundary, and in some sedimentary valleys where deep shallow layers of soft soil can play an important role in earthquake hazard.

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