A study of the relation between ocean storms and the Earth’s hum

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We previously showed that the Earth’s “hum” is generated primarily in the northern oceans during the northern hemisphere winter and in the southern oceans during the summer. To gain further insight into the process that converts ocean storm energy into elastic energy through coupling of ocean waves with the seafloor, we here investigate a 4-day-long time window in the year 2000 that is free of large earthquakes but contains two large “hum” events. From a comparison of the time functions of two events and their relative arrival times at the two arrays in California and Japan, we infer that the generation of the “hum” events occurs close to shore and comprises three elements: (1) short-period ocean waves interact nonlinearly to produce infragravity waves as the storm reaches the coast of North America; (2) infragravity waves interact with the seafloor locally to generate long-period Rayleigh waves; and (3) some free infragravity wave energy radiates out into the open ocean, propagates across the north Pacific basin, and couples to the seafloor when it reaches distant coasts northeast of Japan. We also compare the yearly fluctuations in the amplitudes observed on the two arrays in the low-frequency “hum” band (specifically at 240 s) and in the microseismic band (2–25 s). During the winter, strong correlation between the amplitude fluctuations in the “hum” and microseismic bands at BDSN is consistent with a common generation mechanism of both types of seismic noise from nonlinear interaction of ocean waves near the west coast of North America.

Components: 8173 words, 18 figures, 1 table.
Keywords: Hum of the Earth; microseisms; infragravity waves.
Index Terms: 4546 Oceanography: Physical: Nearshore processes; 4560 Oceanography: Physical: Surface waves and tides (1222); 7255 Seismology: Surface waves and free oscillations.

Received 9 February 2006; Revised 19 June 2006; Accepted 10 July 2006; Published XX Month 2006.


1. Introduction

Since the discovery of the Earth’s “hum” [Nawa et al., 1998], seismologists have tried to determine the source of the continuous background free oscillations observed in low-frequency seismic spectra in the absence of earthquakes.

In the last decade, some key features of these background oscillations have been documented. First, their source needs to be close to the Earth’s surface, because the fundamental mode is preferentially excited [Nawa et al., 1998; Suda et al., 1998] and no clear evidence for higher mode excitation has yet been found. Second, these oscillations must be related to atmospheric processes,
because annual [Nishida et al., 2000] and seasonal
[Tanimoto and Um, 1999; Ekström, 2001] varia-
tions in their amplitudes have been documented.
Finally, they are not related to local atmospheric
variations above a given seismic station, since
correcting for the local barometric pressure flux-
tuations brings out the free oscillation signal in
the seismic data more strongly [e.g., Roul and
Crawford, 2000].

[5] Early studies proposed that the “hum” could be
due to turbulent atmospheric motions and showed
that such a process could explain the corresponding
energy level, equivalent to a M 5.8–6.0 earthquake
every day [Tanimoto and Um, 1999; Ekström,
2001]. However, no observations of atmospheric
convection at this scale are available to confirm this
hypothesis. In the meantime, it was suggested that
the oceans could play a role [Watada and Masters,
2001; Rhie and Romanowicz, 2003; Tanimoto
2003].

[6] Until recently, most studies of the “hum” have
considered stacks of low-frequency spectra for
days “free of large earthquakes. In order to gain
resolution in time and space and determine whether
the sources are distributed uniformly around the
globe, as implied by the atmospheric turbulence
model [e.g., Nishida and Kobayashi, 1999], or else
have their origin in the oceans, it is necessary to
take a time domain, propagating wave approach.
In a recent study, using an array stacking method
applied to two regional arrays of seismic stations
equipped with very broadband STS-1 seismome-
ters [Wielandt and Strecker, 1982; Wielandt and
Steim, 1986], we showed that the sources of the
"hum" are primarily located in the northern Pacific
Ocean and in the southern oceans during the
northern hemisphere winter and summer, respec-
tively [Rhie and Romanowicz, 2004 (hereafter
referred to as RR04)], following the seasonal
variations in maximum significant wave heights
over the globe, which switch from northern to
southern oceans between winter and summer. We
suggested that the generation of the hum involved a
three stage atmosphere/ocean/seafloor coupling
process: (1) conversion of atmospheric storm en-
ergy into short-period ocean waves, (2) nonlinear
interaction of ocean waves producing longer-
period, infragravity waves, and (3) coupling of
infragravity waves to the seafloor, through a pro-
cess involving irregularities in the ocean floor
topography. However, the resolution of our study
did not allow us to more specifically determine
whether the generation of seismic waves occurred
in the middle of ocean basins or close to shore [e.g.,
Webb et al., 1991; Webb, 1998]. The preferential
location of the sources of the Earth's “hum” in the
oceans has now been confirmed independently
[Nishida and Fukao, 2004; Ekström and Ekström,
2005]. In a recent study, Tanimoto [2005] showed
that the characteristic shape and level of the low-
frequency background noise spectrum could be
reproduced if the generation process involved the
action of ocean infragravity waves on the ocean
floor, and suggested that typically, the area involved
in the coupling to the ocean floor need not be larger
than about 100 × 100 km². However, the linear
process proposed may not be physically plausible,
because of the difference in wavelength between
infragravity and elastic waves (S. Webb, personal
communication, 2006).

[6] On the other hand, oceanographers have long
studied the relation between infragravity waves and
swell. Early studies have documented strong cor-
relation between their energy levels, which indicate
that infragravity waves are driven by swell [e.g.,
Munk, 1949; Tucker, 1950]. Theoretical studies
have demonstrated that infragravity waves are
second order forced waves excited by nonlinear
difference frequency interactions of pairs of swell
components [Hasselmann, 1962; Longuet-Higgins
and Stewart, 1962]. A question that generated
some debate was whether the observed infragravity
waves away from the coast are “forced” waves
bound to the short carrier ocean surface waves and
traveling with their group velocity, or “free” waves
released in the surf zone and subsequently reflected
from the beach which, under certain conditions,
may radiate into deep ocean basins [e.g., Sutton
et al., 1965; Webb et al., 1991; Okihiro et al., 1992;
Herbers et al., 1994, 1995a, 1995b]. In particular,
Webb et al. [1991] found that infragravity wave
energy observed on the seafloor away from the
coast was correlated not with the local swell wave
energy but with swell energy averaged over all
coastlines within the line of sight of their experi-
mental sites, in the north Atlantic Ocean and off
shore southern California.

[7] Recently, we analyzed the relation between
ocean storms off-shore California and infragravity
wave noise observed on several broadband seafloor
stations in California and Oregon [Dolenc et al.,
2005a]. We also found that the seismic noise
observed in the infragravity wave band correlates
with significant wave height as recorded on re-
gional ocean buoys, and marks the passage of the
storms over the buoy which is closest to the shore.
More recent results based on data from an ocean floor station further away from shore [Dolenc et al., 2005b] indicate that the increase in amplitude in the infragravity frequency band (50–200 s) associated with the passage of a storm occurs when the storm reaches the near coastal buoys, and not earlier, when the storm passes over the seismic station. This implies that pressure fluctuations in the ocean during the passage of the storm above the

Figure 1. (a) Maximum stack amplitude (MSA, see definition in text) filtered using a Gaussian filter with center period of 100 s for BDSN (blue) and F-net (red), normalized by minimum value of MSA for a given time window. The linear scale on the y axis to the right is used for the normalized MSA to more clearly see the variations in amplitude of the background noise. Black dots represent earthquakes which occurred during the period considered. The corresponding scale is logarithmic (moment magnitude) and is given on the y axis to the left. (b–d) Same as Figure 1a for 150, 200, and 240 s.
station can be ruled out as the direct cause of the low-frequency seismic noise.

[s] In this paper, we investigate these processes further in an attempt to better understand where the coupling between infragravity waves and ocean floor occurs, generating the seismic “hum.” In particular, we describe in detail observations made on event magnitude. This significantly limited the number of usable days in a given year. For example, only 64 days of “earthquake free” data were kept for the year 2000. Among these, we identified the time interval 2000.030 to 2000.034 (i.e., 30 January to 3 February) as a particularly long interval free of large earthquakes, during which the background noise amplitude was unusually high, and during which two large noise events were observed, that could be studied in more detail.

[10] As described in RR04, we considered data at two regional arrays of very broadband seismometers, BDSN (Berkeley Digital Seismic Network) in California, and F-net in Japan. For each array, we stacked narrow-band filtered time domain vertical component seismograms according to the dispersion and attenuation of Rayleigh waves, assuming plane wave propagation from an arbitrary azimuth. Here, we apply a 6 hour running average with a 6 hour running average with a...
time step of 1 hour to the stacked data at BDSN and F-net respectively. At each time step, the stack amplitude has a maximum corresponding to a particular back-azimuth. We consider the resulting maximum stack amplitudes (MSA) as a function of time. We show in Appendix A that the level of the background “hum” is consistent with previous estimates [e.g., Tanimoto and Um, 1999; Ekström, 2001].

We consider the 15 day period 2000.25 to 2000.40. In Figure 1, we plot the MSA as a function of time in four different period bands. We here use a linear scale for the MSA (different from Figure A1) to more clearly see the variations in amplitude of the background noise. We note the well defined signature of large earthquakes, which have a sharp onset, a slower decay and a relatively sharp end. At the time resolution considered here, this onset is practically coincident at both arrays.

Figure 2. (a) Mean stack amplitude averaged over 6 hour sliding window (window shifted 1 hour between resolutions) as a function of time and back azimuth for F-net. A Gaussian filter with center period of 150 s was applied before stacking. (b) Same as Figure 2a for BDSN. (c) Same as Figure 2a for center period of 240 s. (d) Same as Figure 2c for BDSN. The time difference between corresponding energy arrivals at the two arrays on day 31 is about 8–10 hours, with F-net lagging behind BDSN. This is more clearly seen in the shorter-period plot.
The duration of the earthquake signal increases with the size of the earthquake and is typically on the order of 0.5 day for $M_w 6$ and 1–1.5 day for $M_w 7$ earthquakes, after which the signal drops below the average background noise level. This is consistent with what one expects from the decay of Earth circling mantle Rayleigh waves generated by large earthquakes.

Table 1 lists all earthquakes larger than $M_5.0$ during these 15 days, as reported in the NEIC catalog. During the time interval 2000.031 to 2000.034, there are no earthquakes larger than $M_5.5$, yet the background noise rises well above the noise floor, forming two particularly long events, with a very different signature from that of earthquakes: the rise time is longer, the decay very slow and the ratio of the duration of each event to its maximum amplitude, significantly larger. These two noise events are observed on both arrays (i.e., in California and in Japan), and there is a lag time of 3 days.

Figure 3. Same as Figure 1 for days from 340 to 355 in 2002.
several hours between the two arrays. The second event is weaker at the longest periods. We verify that the back-azimuth corresponding to the maximum amplitude is very stable during these two events, as illustrated in Figure 2, which also emphasizes the delay of about 8–10 hours between the main energy arrivals at BDSN and F-net. We will discuss these events in detail in what follows.

This particular “earthquake free window” is unique in that it lasts several days, and the noise events are large. However, noise events with similar characteristics are observed at other times as well. For example, Figure 3 shows a similar plot for the time period 2002.340 to 2002.355, in which we observe a noise event beginning on day 2002.349, showing similar time evolution as for the events in 2000 described above: a slow rise time and lag of ~8–10 hours between the two arrays. It is followed by a second noise event of similar characteristics, but partially hidden behind an earthquake of $M_w > 6$. In what follows, we return to the time period 2000.031–034 for further analysis.

Figure 4. (a) Results of grid search method to locate the source of continuous long-period Rayleigh waves on 31 January 2000. Six hour waveforms Gaussian filtered with center period of 100 s from F-net, BDSN, and 10 European stations are used. Color indicates the mean stack amplitude over a 6 hour time window (2000.031,14:00–2000.031,20:00 UTC) after correcting waveforms at individual stations for attenuation and dispersion. (b) Same as Figure 4a for 150 s.
The large noise events observed on days 2000.031 and 2000.033 after applying a smoothing moving average to the MSA, are the coalescence of multiple smaller events, which, as we showed previously, propagate across the two arrays with the dispersion characteristics of Rayleigh waves (see Figure 1 in RR04). In order to locate the sources of these disturbances in RR04, we applied a back-projection grid-search method to the original time series, after band-pass filtering between 150–500 s, over a 6 hour period containing the maximum stack amplitude on day 2000.031. We showed that the sources of Rayleigh waves that best fit the amplitudes observed both at BDSN and F-net are located in the North Pacific Ocean basin. We here apply the same back-projection method, but using a narrow band filter centered at 150 and 100 s respectively, and obtain a band of source locations which follows the north Pacific shoreline, as illustrated in Figure 4. This is particularly clear at 100 s.

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[15] In order to obtain a stable solution using the grid search method, it is necessary to process a time interval of length about 6 hours, indicating that many of the small events which compose the...
larger noise event on day 2000.031 are too small to be studied individually, at least with this method: the minimum duration of the time window necessary to obtain a stable result is controlled by the available signal/noise ratio in the stacks. Here we show that the large “composite” event, obtained when using the 6 hour moving average, propagates west to east across the whole North American continent, with an amplitude decay consistent with the propagation of Rayleigh waves. Instead of stacking the noise data over an array of broadband seismic stations, we here consider five quiet broadband stations in North America, and, for each of them, we compute power spectral density (PSD) as a function of time, with a 6 hour moving window and a 1 hour lag (Figure 5). All five stations show an increase in background noise during days 31–32, and another one, with smaller amplitude and narrower frequency range, on day 33 (except for CCM for which data are not available on that day). Figure 6a compares the mean Fourier amplitudes at stations CMB, TUC, ANMO and HRV, averaged over the period range 100–200 s and time range 2000.31,00:00 and 2000.32,06:00. We chose this period range, because at lower frequencies, the background noise is dominated by site effects at some of the stations. From the amplitude decay it is possible to obtain a very rough
estimate of the location of the source of Rayleigh waves by forward amplitude modeling. The results are shown in Figure 6c, using CMB, TUC, ANMO and HRV. Because the available azimuth range is not very wide, there is a large uncertainty in the longitude of the inferred source. However, it is compatible with a location near the west coast of North America. Figure 6b compares the observed and predicted average Fourier amplitudes at four stations. The amplitudes are normalized to those of the most western station (CMB) and the predicted amplitudes are computed assuming the Q model of PREM for Rayleigh waves [Dziewonski and Anderson, 1981] and accounting for geometric spreading.

3. Correlation With Ocean Buoy Data

To further investigate the origin of the noise events on days 2000.031 and 2000.033, we now...
Figure 8. Significant wave height map on 31 January 2000 for the north Pacific Ocean, based on WAVEWATCH III. (a–d) Different time windows from 0 to 18 hour with 6 hour interval.

Turn to a comparison with ocean buoy data. We collected significant wave height (SWH) data measured at buoys deployed in the north Pacific by the National Ocean and Atmospheric Administration (NOAA) and the Japan Meteorological Agency (JMA) and operational during those days. We compare SWH time series for this time period to the time evolution of the maximum stack amplitudes at BDSN and F-net for the same time interval. Figure 7 shows such a comparison for buoys located near Japan and near the California coast. The time series on all buoys closely resemble the “source signature” on the seismic stacks, shown here at a period of 240 s. This is the case for the event on 2000.031 as well as for the smaller one on 2000.033. The seismic noise events on BDSN lag those observed on buoy 46059 by about 10–12 hours, but are more or less coincident (to within 1 hour, which is the minimum resolution of these plots) with the events observed on the near shore buoys, indicating that the location of the coupling between ocean waves and the seafloor occurs somewhere between buoy 46059 and the shore, which is consistent with the results of Figure 6c. The ocean storm which generated the short-period waves observed on buoys both near Japan and near the western US moved from east to west across the north Pacific basin. Unfortunately, we could not find any buoy data closer to the eastern coast of Japan, or in other parts of the western Pacific Ocean. To further investigate the source of these waves, we therefore turn to wave models. Figure 8 shows snapshots of the evolution of wave height in the northern Pacific for day 2000.031, from the WAVEWATCH III model [Tolman, 1999]. During that day, a large storm arrives from the west toward the coast of California and Oregon. It reaches the coast, according to the model, between 6h and 12h on day 2000.031. It is followed by a smaller “tail,” about 3000 km behind, which, in turn, according to the WAVEWATCH III model, reaches the coast between 0h and 6h on day 2000.033 (not shown). The following storm system, which forms in the western part of the north Pacific (around longitude 160°E on Figure 8) on day 2000.031 develops into a stronger storm over the next few days. This can be seen in the animation provided by NOAA at http://ursus-marinus.ncep.noaa.gov/history/waves/nww3.hs.anim.200001.gif and http://ursus-marinus.ncep.noaa.gov/history/waves/nww3.hs.anim.200002.gif. This storm is not associated with any significantly increased seismic noise on BDSN or F-net (Figure 1). Notably however, in contrast to the previous one, this storm does not reach the California coast, but dissipates in the middle of the ocean. The distribution of wave heights on Figure 8, together with the observation of the significant delay in the stack energy at F-net with respect to the BDSN (see also Figure 2), leads us to propose the following sequence of events. [17] On day 2000.031, a large storm, which developed two days earlier in the middle of the north
Pacific basin (according to wave models and buoy data) and moving eastward toward North America, reaches the vicinity of the western United States coast. A second storm, weaker, but with similar characteristics, follows by about 2 days. The seismic background noise observed on the BDSN and F-net arrays has the same amplitude signature, as a function of time, as the storms. The process that converts the storm energy into seismic energy, which then propagates as Rayleigh waves, in particular through the North American continent, appears to involve several steps (Figure 9): when the storm approaches the US coast with its rough seafloor topography, short-period ocean waves interact nonlinearly to produce infragravity waves. Part of the infragravity wave energy then converts to seismic waves locally, to produce the background noise event observed on BDSN, and part is reflected back out to the ocean and travels across the Pacific basin, in agreement with oceanographic studies of the generation of infragravity waves [e.g., Munk et al., 1964; Elgar et al., 1992; Herbers et al., 1995a, 1995b]. We estimate that, at ~220 m/s, “free” infragravity waves propagate about 6000–8000 km in 8–10 hours. Consistent with the back-azimuth of the maximum arrival of energy, the conversion from infragravity waves to seismic waves detected on F-net primarily occurs in the vicinity of the western Aleutian arc. We note that the absolute level of MSA is larger at BDSN than at F-net (e.g., Figures 2 and 7), in agreement with the inference that the source for F-net should be comparatively more distant and also weaker.

We infer that free infragravity waves play a role in generating the seismic disturbances in Japan because of the 8–10 hour time delay between stack maxima on BDSN and F-net. This is consistent with observations of remotely generated infragravity waves [e.g., Herbers et al., 1995a]. This time delay is too short for propagation of short-period ocean waves (and also “bound” infragravity waves) from the center of the north Pacific basin, and much too long for propagation of seismic waves from a source near the US coast to Japan. An alternative scenario for the sources of seismic noise on F-net could involve the storm which forms on the Japan side of the Pacific in the middle of day 2000.031 (Figure 8). However, we rule this out, because this storm intensifies only later and reaches its peak around 03h on day 2000.032, which is much later than the long-period seismic peak on F-net.

We note that the generation of large infragravity waves from short-period ocean waves along the east coast of the Pacific (Canada, US) rather than the west coast (Japan) is due to the fact that prevailing winds are westerlies, and therefore most ocean waves are driven from the west to the east, interacting nonlinearly only with the coasts on the east side of ocean basins.
In summary, the seismic sources that form the composite events on days 2000.031 and 2000.033 are distributed around the Pacific, both in time and space, but have a common cause: a strong storm system which “hits” the North American coast broadside. A similar type of storm which reaches North America from the west, occurs on day 2002.349 (Figure 10), causing the disturbances observed on Figure 3. We infer that the efficiency of generation of seismic waves is particularly high for these storms, due to their direction of approach to the North American coast, and the fact that these storms actually reach the coast. This is why we can observe these remarkable “noise events” on the stacks at BDSN and F-net so clearly. We have evidence of directionality of the process, in that station COL (Alaska) does not show any increase of seismic noise in the 70–250 s pass band during the same time period. At least, it is below detection level by our methodology involving PSD spectra, even though there is an indication, from the noise in the microseismic bandpass (2–25 s) of a storm reaching the Alaska coast nearby at the end of day 2000.031 (Figure 11). Such directionality would also explain why we can so clearly follow the particular seismic disturbance on day 2000.031 across North America.

Other north Pacific storms must also generate long-period seismic noise, however, the corresponding noise “events” cannot often be identified as clearly because they are either hidden behind large seismic events, or do not have sufficient amplitude levels to rise above the average noise level on the two seismic arrays considered. We note that many winter storms never reach the north American coast, or turn further north into the Gulf of Alaska. A systematic analysis of storm characteristics in the north Pacific in relation to the “hum” is beyond the scope of this paper and will be addressed in a further study.

4. Comparison With Microseisms

We have shown that infragravity waves generated by winter storms in the north Pacific Ocean contribute to the source of the low-frequency “hum” events observed in California and Japan.

The nonlinear wave interactions that give rise to infragravity waves are also responsible for the generation of double-frequency microseisms [e.g., Hasselmann, 1962, 1963; Longuet-Higgins, 1950], which are themselves correlated with the wind wave spectrum [e.g., Babcock et al., 1994; Webb and Cox, 1986; Bromirski and Duennebier, 2002], and are known to be generated primarily locally near the coast [e.g., Haubrich and McCamy, 1969; Webb, 1998; Bromirski et al., 2005].

Therefore we next investigate the relationship between microseisms, ocean storms and the low frequency “hum.” Even though there are two types of microseisms, primary (at periods lower than 10 s) and secondary, or “double-frequency,” at periods...
around 6–8 s [e.g., Friedrich et al., 1998], and their generation mechanisms are different [e.g., Hasselmann, 1963; Webb, 1998], the double-frequency microseisms dominate the spectra and we will only consider those in the discussion that follows.

We first computed mean Fourier amplitudes in the microseismic period band (2 to 25 s) at individual stations of BDSN and F-net for the time interval 2000.031–2000.035. We used moving windows of duration 30 min, shifted by 10 min. We removed mean and trend before computing Fourier amplitudes. We then compared them to

Figure 11. (a) Power spectral density (PSD) at COLA in Alaska. (b) Mean Fourier amplitude in the period range 2–25 s for COLA.
near-by buoy data (Figures 12 and 13). Along the coast of California (Figure 12), the mean seismic amplitudes show the signatures of the two noise events already discussed at low frequency (on days 2000.031 and 2000.033), which are also well defined on the buoy data. The eastward moving storm arrives first in northern and central California, as seen by the slight delay in its wave height signature at buoy 46062 compared to the other two buoys (see also the data from buoy 46059 on Figure 7). The timing of the peak of microseismic noise at the three stations and the fact that the amplitude at station ISA is smaller by about a factor of 3 than at BKS, indicate that the generation of the microseisms occurs closer to the central and northern California buoys.

Unfortunately, only data for three buoys are available around Japan for this time period. However, we note that the mean microseismic Fourier amplitudes at the three seismic broadband stations closest to the buoys show a good correlation with SWH data (Figure 13). We also note that, contrary to the observations in California, the timing and shape of the microseismic amplitude variations is different from that at “hum” frequencies, and

Figure 12. (a) Location map of BDSN and TerraScope stations (black triangle and blue squares) and buoys (green dots). Blue squares are seismic stations closest to the corresponding buoys. (b–d) Significant wave heights measured at buoy 46027, 46026, and 46052, respectively. (e–g) Mean Fourier amplitude (count/Hz) over the period range 2–25 s for YBH, BKS, and ISA.
varies significantly with location of the station in the array (Figure 14), indicating that, in Japan, the sources of the microseismic noise and of the hum are distinct: the low-frequency noise is related to that observed on the eastern side of the Pacific (with a delay which we attribute to the propagation of infragravity waves across part of the Pacific Ocean), whereas the microseismic noise maximum occurs significantly earlier (on day 2000.030). In fact, the time histories of microseism energy at stations within Japan differ and presumably depend on the location of each site relative to each storm track.

[27] To further investigate the relation between microseismic noise and the low-frequency hum, we need to be able to compare amplitude levels in the two frequency bands for long time intervals (e.g., a whole year). To do so effectively, we developed a data processing method that avoids...
eliminating the numerous time windows that are contaminated by earthquakes.

Removing the effect of earthquakes at low frequencies is difficult to do precisely, due to the presence of lateral heterogeneity in the Earth and the relatively low attenuation. In order to minimize their effects, we compute the minimum value, as a function of time, in a sliding 1.5 day interval, of the scaled MSA time series, using a moving time window with a 6 hour shift. This effectively removes some large amplitude peaks due to earthquakes.

Figure 14. (top) Location map of seismic stations grouped by their locations in F-net. Different colors indicate different grouping. (bottom) Mean Fourier amplitude (count/Hz) over the period range 2–25 s for two selected stations in each group shown on the map at the top in black (left column), in blue (center column), and in red (right column). The variations in amplitude for all five stations (three of them are not shown) in the same group show similar overall trends.
580 quakes, but not all. We then apply a low-pass filter
581 with a corner period of 1 day to the time series
582 obtained in the previous step. This further removes
583 most of the earthquake-related peaks, except for
584 those with the longest duration, corresponding to
585 the largest earthquakes (Figures 15a, 15b, and 15c).
586 We also compute the mean Fourier amplitude in the
587 microseism band (2–25 s) for seven BDSN
588 stations. Here the contamination by large earth-
589 quakes is not as severe and we only remove those
590 points which correspond to large temporal gra-
591 dients. To do so, we empirically determined a
592 gradient threshold between two consecutive points
593 in the amplitude time series: if the measured
594 gradient is higher than the threshold, we remove
595 the end point and test the gradient value for
596 successive end points, until the gradient drops
597 below the threshold. Finally, we low-pass filter
598 the amplitude time series with a corner period of
599 1 day (Figures 15d, 15e, and 15f). This effectively
600 removes most of the earthquake signals.

[29] We compare the filtered “hum” and micro-
601 seism amplitude time series over a period of one
602 year, for each array. In the case of California
603 (BDSN), the level of low-frequency noise does
604 not vary systematically with time (Figure 16a), but
605 there is a seasonal variation in the microseismic
606 amplitude, with a minimum during northern hemi-
607 -sphere summer time, as is also seen in the ocean
608 wave height data (Figure 16b). This indicates that
609 the sources of energy for the long-period and short-
610 period noise are different during the summer. The
611 variation in microseismic amplitude at BDSN sta-
612 tions is clearly related with ocean wave height
613 measured by local buoys (Figure 16b). We can
614 see a similar trend for F-net, but the correlation of
615 the variation in short-period amplitudes and ocean
616

Figure 15. (a) Scaled long-period MSA, Gaussian filtered with center period of 240 s (blue curve) for BDSN. Circles indicate earthquakes. (b) Black curve: same as in Figure 15a. Green curve: minimum obtained after applying moving time window with duration of 1.5 days and 6 hour shift. (c) Green curve: same as in Figure 15b. Blue curve: after low-pass filtering with corner period of 1 day. (d) Mean Fourier amplitude over the microseismic band (2–25 s) averaged over 7 BDSN stations (black curve). Dots are earthquakes as in Figure 15a. (e) Black curve: same as in Figure 15d. Green curve: after removing large gradient peaks. (f) Green curve: same as in Figure 15e. Red curve: low-pass filtered with corner period of 1 day.
wave data is weaker than in the case of BDSN (Figures 16c and 16d).

[30] Removing the time periods contaminated by the largest events, and restricting our analysis to northern hemisphere winter (January to March and October to December), we compute the correlation coefficients between the low-frequency and high-frequency filtered noise time series, for three consecutive years, at BDSN. Correlation coefficients are significant, between 0.39 and 0.60 (Figure 17). This indicates that in the winter, both the low-frequency hum and the microseismic noise observed at BDSN are generated locally. On the other hand, the corresponding correlation coefficients for F-net are generally much lower: for the first 3 months of each year, respectively: 0.11, (N/A), 0.21; for the last 3 months of each year, respectively: 0.22, 0.02, 0.30.

[31] The correlation between the hum and microseismic noise at BDSN during the winter is compatible with a common generation mechanism for both types of seismic noise, involving nonlinear interactions between surface ocean waves giving rise, on the one hand, to double-frequency microseisms, and on the other, to infragravity waves [e.g., Hasselmann, 1962]. The fact that the correlation is somewhat weaker at F-net is in agreement with our proposed scenario, in which the dominant effect is that of storms moving from West to East across the Pacific and reaching the west coast of North America to produce low-frequency seismic “hum.”

5. Conclusions

[32] We have made progress in clarifying the mechanism of generation of continuous free oscillations, based on the observations for a time interval free of earthquakes during which two large long-period noise events are present in the MSA at BDSN and F-net. We have shown that these events can be related to a particular winter storm system.

[33] A perturbation in the atmosphere, typically a winter storm moving eastward across the north Pacific basin, generates short-period ocean waves. As the storm reaches the north-American coast, the nonlinear interaction between ocean waves generates long-period infragravity waves, some of which convert locally to long-period seismic energy, and others propagate long distance across the ocean basin and couple to the seafloor near northeastern coasts. The resulting long-period seismic waves propagate over the globe and give rise to the “hum.” In particular, we were able to track the seismic energy generated off-shore California by the storm considered on day 2000.031, throughout the North American continent.
The directionality of the “hum” radiation suggested by our data needs to be further characterized, in particular for the benefit of studies of structure based on the analysis of noise cross-correlations [e.g., Shapiro et al., 2005], at least at low frequencies. Indeed, the sources of low-frequency seismic noise can no longer be considered as uniformly distributed either in time, or in space.

The annual fluctuations of long- (hum band) and short- (microseism band) period seismic amplitudes at BDSN and F-net show quite different features. We can clearly see the seasonal change in amplitude in the microseism band (2–25 s) with a minimum during northern hemisphere summer, whereas the amplitude in the hum band (here considered at ~240 s) does not show clear seasonal variations. We also observed a significant correlation between seismic amplitudes at BDSN in the microseism and hum bands during northern hemisphere winter. We had previously documented that the source of the hum observed at BDSN and F-net shifts from the northern Pacific to the southern oceans between winter and summer, so that the...
sources are more “local” in the winter than in the summer. In contrast, microseisms propagate less efficiently at large distances, so the source is primarily local. These observations are in agreement with a common mechanism for the simultaneous generation of short- and long-period seismic noise near the California coast, as inferred from theoretical studies.

Appendix A

[36] We estimate the background level of the low-frequency seismic energy by determining a scaling...
factor between the observed peak amplitudes (from
the MSA at 240 s), and the moment magnitudes of
the corresponding earthquakes (\(M_w > 6.0\)), as listed
in the Harvard CMT catalog \cite{Dziewonski and
Woodhouse, 1983}.

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Acknowledgments

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We wish to thank the staff of BDSN, F-net, JMA, NOAA
for providing high-quality and continuous seismic and ocean
buoy data. We thank Sphar Webb and an anonymous reviewer
for helpful comments. Berkeley Seismological Laboratory
contribution 06-XYZ.

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