GEOPHYSICAL RESEARCH LETTERS, VOL. 25, NO. 23, PAGES 4257-4260, DECEMBER 1, 1998

Theoretical modeling of seismic noise propagation in firn at the South Pole, Antarctica

Donald G. Albert

U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, NH 03755

Abstract The problem of interfering noise (produced by ground vehicles) on teleseismic arrivals recorded by Global Seismic Network sensors at South Pole Station is addressed. Using the wavenumber integration method, theoretically calculated seismograms show that installing the GSN sensors in a borehole 200 to 300 m deep, 10 km away from the Station, will significantly reduce the vehicle-generated noise and improve signal quality. Because the intrinsic attenuation of seismic waves propagating in the polar firn is low, most of the predicted noise reduction results from wavefront spreading, Rayleigh wave amplitude decay with depth, and from placing the sensors below the refractive waveguide that traps much of the seismic energy in the near surface layers.

Introduction

For many years, Incorporated Research Institutions for Seismology (IRIS) and the Albuquerque Seismological Laboratory of the U.S. Geological Survey have been operating a seismological station (designated SPA) at the South Pole as part of the Global Seismological Network (GSN). SPA is of critical importance because few other stations operate in Antarctica. Unfortunately, increasing operational activities at the South Pole Station (vibrations caused by generators, heavy vehicle movement, and aircraft landings and taxiing) are beginning to interfere with and contaminate recordings made by SPA. To reduce the effects of this cultural noise, the SPA sensors (Streckeisen Mdl STS-1/VBB 3 component), installed in a 4-m-deep vault, will be replaced by new borehole sensors some distance away from the Station. The highest recorded frequency also will be increased from the current 10 Hz to 32 Hz. The question that remains is how far away, and how deep, should the seismograph sensors be to reduce the noise to acceptable levels. Cost and the difficulty of maintaining continuous year-round operation in this harsh environment make it desirable to emplace the new sensors as close to the Station as possible, while keeping the noise below acceptable levels.

One of the main physical features of the Antarctic polar ice sheet is the steep near-surface seismic velocity gradient, caused by firn densification. This gradient forms a waveguide that traps much of the energy from seismic sources at the surface (which are considered to be *noise* in this paper), so placing the SPA sensors below this level should greatly reduce surface-generated noise and enhance the signal-to-noise ratios for teleseismic signals. However, such gradients make the prediction of noise decay complicated. For the optimum placement of the SPA sensors, a model based on the physical properties of the firn and ice at the South Pole was used to compute transmission loss of seismic noise as a function of distance, depth, and frequency. In addition, synthetic seismograms were constructed to gain understanding of the propagation characteristics of this environment.

Paper number 1998GL900155.

Computational Method

The wavenumber integration method, implemented in a computer code known as OASES [*Jensen et al.*, 1994; *Schmidt*, 1987; *Schmidt*, 1996; *Schmidt and Tango*, 1986], is used to investigate seismic waves propagating in the polar ice sheet. This method treats the polar firn as a series of plane, horizontal, isovelocity viscoelastic layers, and solves the wave equation by applying the usual boundary conditions (At the surface, the normal and tangential stresses are zero, while at the layer boundaries these stresses and the vertical and radial particle velocities are continuous.) using the direct global matrix approach [*Schmidt and Tango*, 1986]. Asymptotic expansions of the involved Bessel functions, valid for large ranges and high frequencies (wavenumber times distance greater than 1), are used to numerically evaluate the solution (an integral over wavenumber) and to determine the response of the medium at a single frequency. Synthetic seismograms are calculated by obtaining the solution at a number of frequencies over the bandwidth of interest, and using an inverse Fourier transform to construct the time domain pulse. A complete, detailed explanation of the method appears in Chapter 4 of *Jensen et al.* [1994].

Properties of Polar Firn at South Pole

As the firn densifies with depth because of overburden pressure, it also increases markedly in stiffness by sintering [*Gow*, 1963a; *Gow*, 1975; *Paterson*, 1994], which produces rapidly increasing seismic velocities with depth. Thus, the upper part of the firn has a strong velocity gradient, with compressional wave speeds increasing from about 500 m s^{-1} at the surface to a maximum of about 3900 m s^{-1} in the solid ice. The shear wave velocity also increases, to a maximum of about 1950 m s^{-1}. The depth where the maximum density of solid ice is reached depends primarily on the temperature; based on ice cores, this depth is probably more than 200 m [*Giovinetto*, 1960; *Gow*, 1965; *Herron and Langway*, 1980; *Kuivinen et al.*, 1982] at the South Pole (mean temperature -51° C). The strong velocity gradient in the near surface firn creates a waveguide, refracting downgoing seismic waves back toward the surface.

The seismic calculations require that compressional wave velocity V_p, shear wave velocity V_s, density ρ, compressional and shear attenuation rates, and layer thicknesses be specified as input. At the South Pole, the only seismic refraction measurement available was an early effort done using analog equipment and wet photographic processing [*Weihaupt*, 1963]. Those results agreed with refraction measurements conducted near the Pole during a traverse [*Robinson*, 1962]. Weihaupt's measured seismic velocities, supplemented with relationships derived from

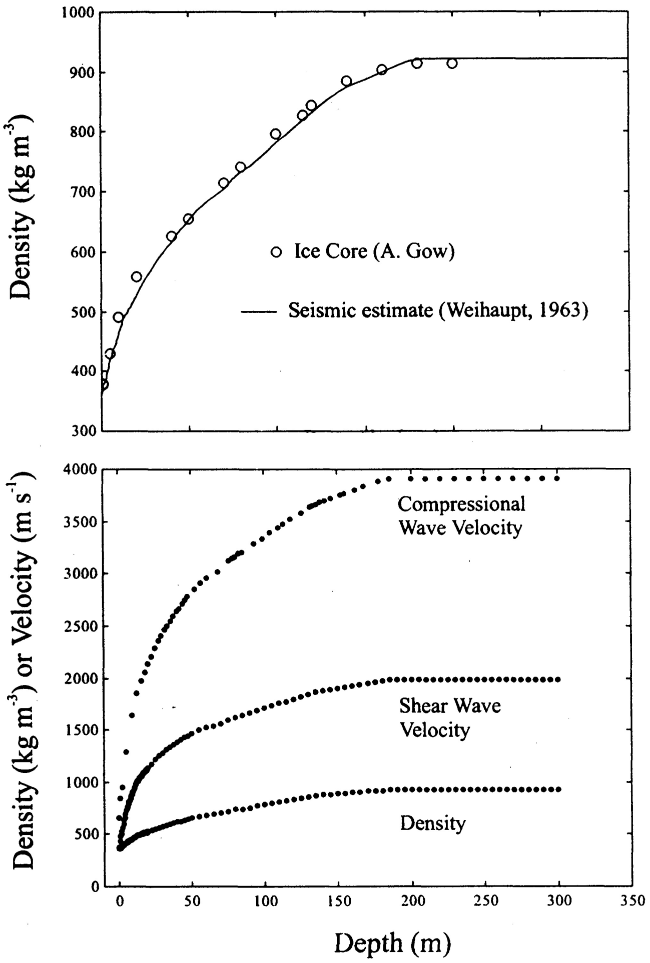

Figure 1. (top) Comparison of firn and ice core density measurements (circles, values courtesy of A. Gow) with density estimated (solid line) using seismic data [*Weihaupt*, 1963] and equation (1). The firn-ice transition occurs at 115 - 120 m. (bottom) Compressional wave velocity measurements at the South Pole [*Weihaupt*, 1963], along with the shear wave velocity and firn and ice density values that were derived from the compressional velocity measurements.

measurements at other locations, were used to construct the model of firn properties at the South Pole according to the following procedure.

First, V_p as a function of depth was determined from the seismic measurements and was discretized at the desired model layer depths by spline or linear interpolation. Below a few hundred meters, the temperature increases slightly [*Paterson*, 1994], causing the seismic velocity to decrease. Next, the density ρ was determined from V_p using the empirical relation [*Kohnen*, 1972]

$$\rho(z) = \rho_{\max}\left[1+\left[\frac{V_{\max}-V_p(z)}{2250}\right]^{1.22}\right]^{-1} \quad (1)$$

where $V_{\max}$ and $\rho_{\max}$ are the maximum compressional wave velocity and ice density at this location. Fig. 1 shows good agreement between densities estimated using Weihaupt's seismic measurements in equation (1) and measurements on firn and ice cores at the South Pole [*Giovinetto*, 1960; *Gow*, 1965; *Kuivinen et al.*, 1982]. The shear wave velocity V_s was computed from the compressional wave velocity at each depth using the relation

$$V_s = \sqrt{\frac{1-2n}{2-2n}}\, V_p\ , \quad (2)$$

where n is Poisson's ratio. From seismic compressional and shear wave velocity measurements on warmer firn on the Ross Ice Shelf [*Albert*, 1978], Poisson's ratio increases from 0.2 at a density of 200 kg m^{-3} to 0.29 at 490 kg m^{-3}, reaching a maximum of 0.327 for ice with densities of 900 kg m^{-3} or greater. (This relationship is quite similar to one determined from refraction measurements at the colder Byrd Station [*Kohnen and Bentley*, 1973]; the velocity gradients controlling the waveguide effect are quite similar for both cases.) The resulting values of V_p, V_s, and ρ, used as input to the calculations, are shown in Fig. 1.

Seismic refraction measurements at Byrd Station [*Bentley and Kohnen*, 1976] were used to determine a value of the quality factor Q for compressional waves of about 715 at 136 Hz, corresponding to $\alpha \cong 0.04$ dB per wavelength. Because the coefficient is so low, intrinsic attenuation in the firn and will be a minor factor in reducing the surface-generated noise. The calculations reported here were done both using this value of compressional wave attenuation and using the assumption of purely elastic propagation without intrinsic attenuation.

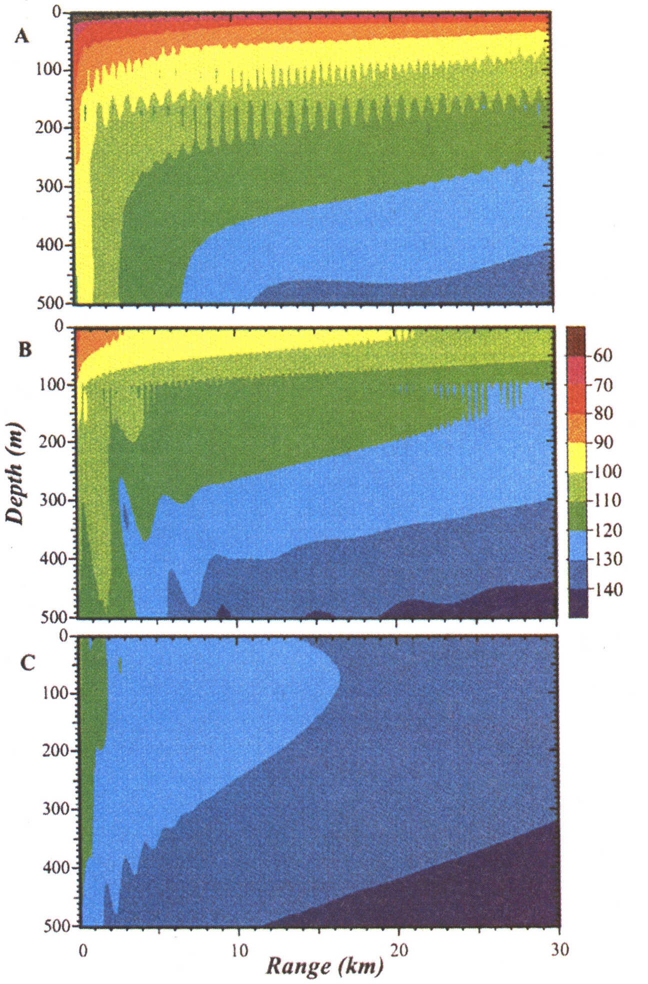

Figure 2. Calculated transmission loss (in dB) for the vertical particle velocity at the South Pole. A vertical point source was located at the surface and the effects of intrinsic attenuation are ignored. (a) 30 Hz, (b) 10 Hz, (c) 3 Hz.

Results

An input model of firn properties with 56 layers was used. For the highest frequency to be recorded by the new sensors, 32 Hz, the wavelength ranges from about 20 m at the surface to 130 m in the solid ice. The input model layers were 1 m thick from the surface to 20 m depth, 5 m thick from 20 to 100 m, and 10 m thick from 100 m to 300 m. A constant velocity layer extending from 300 m to 3 km was underlain by a half space of slightly lower velocity. This model was selected because calculated seismograms for a model with half the layer thickness throughout gave very similar results.

Fig. 2 shows the calculated transmission loss in dB, defined as 20 times the log of the seismic amplitude in m/s (i.e., 0 dB = 1 m/s), for frequencies of 3, 10, and 30 Hz produced by a vertical point source at the surface. The figures show that moving outward along the surface causes very little noise reduction. Once a horizontal range of about 10 km has been reached, significant noise reduction will occur by placing the sensors beneath the firn layers. A depth of 200 – 300 m is very effective in reducing the noise level at 10 Hz and above, but less effective at the lower frequencies because of the larger wavelength.

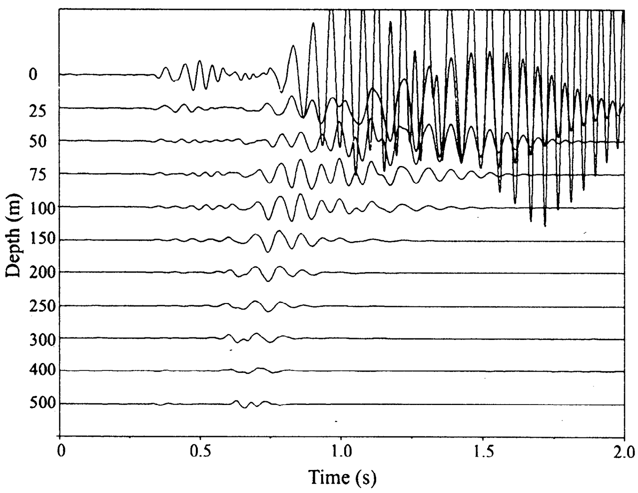

Figure 4. Calculated vertical particle velocity seismograms at various depths for a range of 1 km at South Pole for a vertical point source at the surface. This true amplitude display shows the rapid attenuation of the Rayleigh wave with depth in the firn.

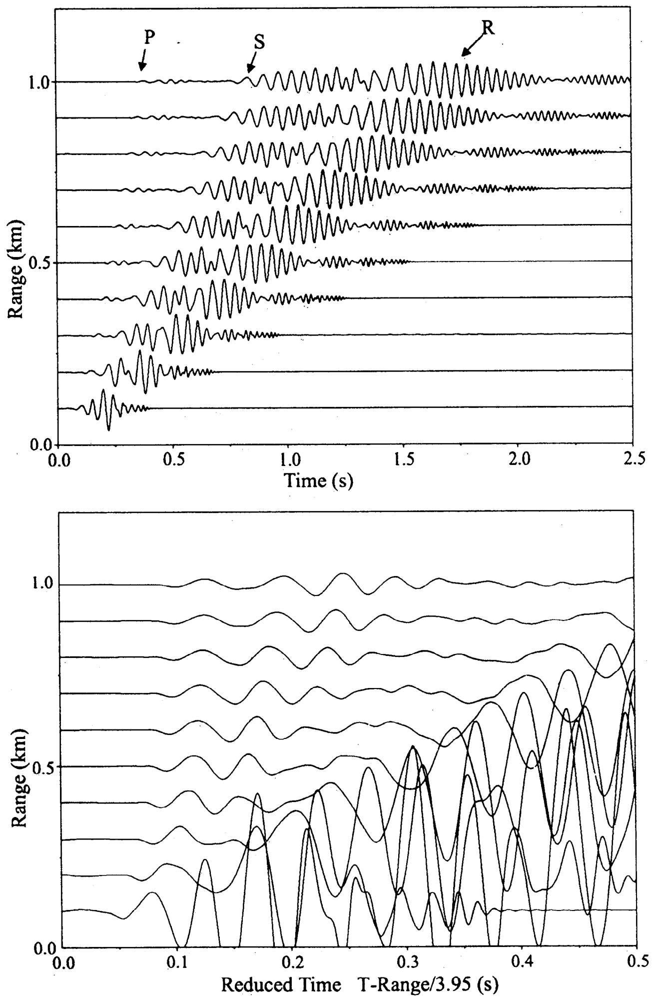

Figure 3. Calculated vertical-particle-velocity seismograms at the surface for South Pole, for a vertical point source at the surface. Intrinsic attenuation was 0.04 dB per wavelength. (a) Complete response shown, with trace amplitudes scaled by range. (b) A shorter time window at higher gain, showing the multiply refracted compressional body wave arrivals, a characteristic feature of the high near-surface velocity gradient

The calculations assume that a source of the selected frequency vibrates for an infinitely long time, and the resultant standing wave pattern is determined. Zones of constructive and destructive interference produce the amplitude lobes shown in Fig. 2, but it should be realized that these zones might not appear for impulsive noise, because the travel times of the direct, reflected, and surface wave arrivals will be different.

Fig. 3 shows surface seismograms calculated from the South Pole model. A bandwidth of 0 to 30 Hz was used in the calculations, and the traces have been multiplied by range, so that cylindrically spreading wavefronts will be shown at a constant amplitude. These waveforms are similar to experimental measurements on polar firn [e.g. *Albert and Bentley*, 1990; *Robin*, 1958; *Robinson*, 1968]. The initial arrival is from the deepest refracted compressional wave. This arrival is followed by a series of multiply refracted compressional waves, and then by shear body waves and the large amplitude Rayleigh wave. This final wave shows dispersive properties as expected, with the lower frequencies arriving earlier than the higher frequencies. Because of its large amplitude, the Rayleigh wave is the main contributor to the near-surface particle velocities and thus to the noise interference problem at the South Pole.

Fig. 4 shows the calculated seismograms at a range of 1 km, for depths of 0 to 500 m. These particle-velocity traces at different depths are plotted at the same gain, so that comparisons from one trace to another show the actual amplitude differences. Since the Rayleigh wave is a surface wave, it is expected to decrease exponentially in amplitude with depth below the surface, and its rapid decay with depth is very clear in the figure. At a depth of 100 m, the maximum amplitude signal is from the seismic body waves. The arrival times identify these as shear waves for the vertical component (Fig. 4) and compressional waves for the horizontal component (not shown).

Concluding Remarks and Recommendations

Theoretical calculations of seismic noise propagation in the polar firn at the South Pole have produced seismograms with realistic appearances and waveform characteristics. Although a direct comparison of the calculated seismograms with measured waveforms is hampered by the lack of digital experimental recordings, a qualitative comparison does show realistic wave arrival times, wave types, and waveform characteristics. The multiply refracted compressional wave arrivals early in the record and the dispersion shown by the large-amplitude Rayleigh waves give support to the impression that the theoretical calculations are complete and accurate.

The computational results indicate that moving the GSN seismometers 10 km away from the South Pole Station and installing them in a borehole at least 200 – 300 m deep should greatly reduce the interference from surface-generated noise and improve the teleseismic signal quality. Because of the low intrinsic attenuation in cold polar firn, most of this noise reduction will be produced by wavefront expansion with distance, by the Rayleigh wave amplitude decay with increasing depth, and by placing the sensors beneath the waveguide in the firn that traps most of the seismic energy originating at the surface.

Although not part of this modeling effort, the SPA sensors also suffer from low frequency noise (below 0.05 Hz) caused by creep within the shallow vault containing the sensors. This low frequency noise is also expected to be reduced since the borehole installation will provide a much more stable platform for these sensors. *Gow* [1963b], for example shows very little borehole deflection over a 4 year period.

Experimental measurements in a shallow borehole to confirm the theoretical modeling results would be prudent, since the predicted Rayleigh wave decay, although theoretically well-founded and commonly observed in other earth materials, has not been experimentally observed in polar firn.

Acknowledgements: I thank Charles R. (Bob) Hutt and his coworkers for providing information about the noise and sensor characteristics at the South Pole, Charles Bentley for useful discussions about Antarctic seismic data, and Anthony Gow for providing firn-density values. Steve Arcone, Anthony Gow, Charles Bentley, and an anonymous reviewer provided helpful technical comments. I also thank Rhett Butler of the Incorporated Research Institutions for Seismology (IRIS), Charles R.Hutt of the U.S. Geological Survey Albuquerque Seismological Laboratory (USGS/ASL), and Scott Borg of NSF for funding this study.

References

Albert, D.G., Seismic investigations on the Ross Ice Shelf, Antarctica, M.A. thesis, University of Texas at Austin, 1978.

Albert, D.G., and C.R. Bentley, Seismic studies on the grid eastern half of the Ross Ice Shelf: RIGGS III and RIGGS IV, in *The Ross Ice Shelf: Glaciology and Geophysics*, edited by C.R. Bentley,and D.E. Hayes, pp. 87-108, American Geophysical Union, Washington, D.C., 1990.

Bentley, C.R., and H. Kohnen, Seismic refraction measurements of internal friction in Antarctic ice, *Journal of Geophysical Research,81*, 1519-1526, 1976.

Giovinetto, M.B., USNC-IGY Antarctic glaciological data - South Pole Station, Ohio State University Research Foundation, Columbus, OH, 1960.

Gow, A.J., Age hardening of snow at the South Pole,*Journal of Glaciology*, *4* (35), 521-536, 1963a.

Gow, A.J., Results of measurements in the 309 meter bore hole at Byrd Station, Antarctica, *Journal of Glaciology*, *4* (36), 771-784, 1963b.

Gow, A.J., On the accumulation and seasonal stratification of snow at the South Pole, *Journal of Glaciology*, *5* (40), 467-477, 1965.

Gow, A.J., Time-temperature dependence of sintering in perennial isothermal snowpacks, *IAHS Publication*, *114*, 25-41, 1975.

Herron, M.M., and C.C. Langway, Firn densification: an empirical model, *Journal of Glaciology*, *25*, 373-385, 1980.

Jensen, F.B., W.A. Kuperman, M.B. Porter, and H. Schmidt, *Computational Ocean Acoustics*, American Institute of Physics, NY, 1994.

Kohnen, H., Uber die Beziehung zwischen seismischen Geschwindigkeiten und der Dichte in Firn und Eis [On the relation between seismic velocities and density in firn and ice], *Z. Geophys.*, *38*, 925-935, 1972.

Kohnen, H., and C.R Bentley, Seismic refraction and reflection measurements at Byrd Station, Antarctica, *Journal of Glaciology, 12(64)*, 101-111, 1973.

Kuivinen, K.C., B.R. Koci, G.W. Holdsworth, and A.J. Gow, South Pole ice core drilling, 1981-1982,*Antarctic Journal of the U.S.*, *17*, 89-91, 1982.

Paterson, W.S.B., *The Physics of Glaciers. Third edition.* Pergammon Press, NY, 1994.

Robin, G.d.Q., Seismic shooting and related investigations, in *Norwegian-British-Swedish Antarctic Expedition, 1949-51*, Norsk Polarinstitutt, Oslo, 1958.

Robinson, E.S., Results of geophysical studies on theMcMurdo to South Pole traverse, University of Wisconsin, Geophysical and Polar Research Center, Madison, WI, 1962.

Robinson, E.S., Seismic wave propagation on a heterogeneous polar ice sheet, *Journal of Geophysical Research*, *73* (2), 739-753, 1968.

Schmidt, H., SAFARI: Seismo-acoustic fast field algorithm for range independent environments. User's guide, SACLANT ASW Research Centre, La Spezia, Italy, 1987.

Schmidt, H., OASES Version 2.0 application and upgrade notes, MIT Dept. of Ocean Engineering, Cambridge, MA, 1996.

Schmidt, H., and G. Tango, Efficient global matrix approach to the computation of synthetic seismograms, *Geophysical Journal of the Royal Astronomical Society*, *84*, 331-359, 1986.

Weihaupt, J.G., Seismic and gravity studies at the South Pole,*Geophysics*, *28*, 582-592, 1963.

D.G. Albert, Geophysical Sciences Division,
USACRREL, 72 Lyme Road, Hanover, NH 03755.
(e-mail: dalbert@crrel.usace.army.mil)

(Received January 13, 1998; revised September 1, 1998;
accepted September 8, 1998.)