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ANNUAL REPORT - AFOSR

THE CENTRAL CALIFORNIAN LARGE-SCALE ARRAY

Contents

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1. General Summary of Contract Research for 1968.

The telemetry network has continued to prove its worth as a seismological observation system. Reasonably precise locations of earthquakes down to 2.5 magnitude which occur locally are routinely and quickly determined at the Stations. This information, together with magnitudes, is used by many institutions for research and other purposes; for example, Berkeley locations of local shocks are now quoted regularly by the USCGS. Significant teleseisms are also located on twenty-four hour a day basis and the information supplied urgently to the Pacific Early Warning System and others. The telemetry system adds greatly to this capability.

Several lines of research have depended heavily on the UC network in the last year. In the first place, properties of S waves from local earthquakes have received special attention. Synthetic seismograms of SH motion have been compared with observed motion from local shocks. Among other conclusions, the results give a means of improving focal depth estimation for crustal shocks.

Secondly, a systematic study of regional tectonics using precise determinations of the seismicity and first-motion P polarities has been completed. The results plainly show a considerable regional uniformity and persistence of patterns of dilatations and compressions which are possessed by small shallow-focus earthquakes in northern California. Domination of the regional stress pattern on earthquake mechanism may be found worldwide and may be important in identifying the source of seismic radiation as an earthquake.

Crustal structure within the area has received further study. Two methods have been used: frequency spectral analysis of P waves from teleseisms and, secondly, the use of seismic refraction methods using earthquakes and blasts.

The variation in Californian crustal structure along transverse profiles from the Pacific Ocean to the Basin Range Province in Nevada has long been a question of wide interest. No really satisfactory seismic profile has ever been obtained which includes stations across the Great Valley of California. Such a profile has now been established in collaboration with the ESSA Earthquake Mechanism Laboratory, using the underground explosions BOXCAR and BENHAM as seismic sources.

We have also started a study of the aftershocks and "collapses" which are recorded on the Berkeley network stations, particularly Jamestown, following atomic detonations in Nevada.

The stations of the telemetry array have continued to yield precise travel-time data for various P and S phases. The year 1968 was not a seismically active one so far as large teleseisms are concerned. However, we have obtained additional seismograms on magnetic tape from distant earthquakes which add key points to the azimuthal anomaly curves for P and S waves. Azimuthal anomalies of over 10° in S from teleseisms have been detected for some directions.

2. Research Reports.

2.1 Seismological Inferences from S Wave Studies Using Local Earthquakes.

The first detailed study of the nature of recorded S waves from local earthquakes in California has been completed. The study arose from our interest in (a) using S observations to improve focal depth estimates and in (b) using S polarization as additional constraints on the focal mechanism of local earthquakes. The main results are set out below.

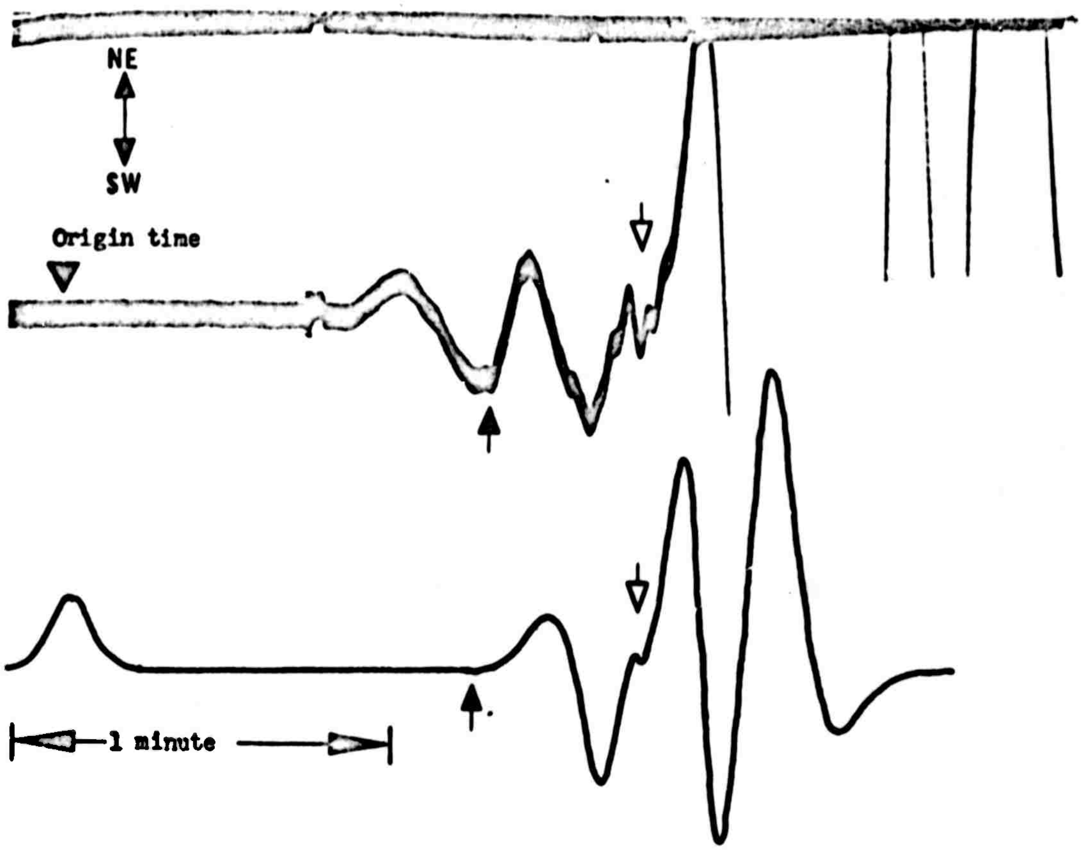
1. The numerical evaluation of the analytical solution obtained using Cagniard's technique for an ideal layer-half space model has yielded close agreement with the general features of the SH motion observed within 1000 km from earthquakes in California.

2. The SH motion recorded at Berkeley (BRK) on short-period instruments from earthquakes within 100 km is a sharp, distinct motion with an apparent surface velocity of about 3.60 $\pm$ .05 km sec. Although the data were not considered sufficient for integration of the travel-time curve, the average crustal shear velocity in the vicinity of BRK should not vary greatly from the apparent surface value. The theory developed predicts this to be the direct arrival, followed by only a few strong reflections at these distances. At the station VIN this arrival is slower (3.51 $\pm$ .06 km/sec) and is followed by an arrival of higher amplitude and frequency with an apparent surface velocity of 3.00 $\pm$ .06 km/sec. Whether this second arrival at VIN is due to a lateral or vertical velocity variation is undetermined. At both VIN and BRK a well-defined initial SH arrival from earthquakes beyond about 100 km was not found on conventional short (1.0 sec) and long (30.0 sec) period instruments. At about 100 km the horizontal component of the diffracted Sn wave should be the initial SH arrival.

3. The SH component of Sn from regional earthquakes ($\Delta > 100$ km) in California is difficult to identify except on records from an instrument of extended long-period response. On such records, the Sn motion is of relatively long period and low amplitude when compared with later reflected arrivals of higher amplitude and frequency. The theoretical seismograms constructed for crustal sources at these distances show this general pattern of motion. The observed Sn arrival has an apparent surface velocity of about $4.5 \pm .2$ km/sec. Close agreement between actual records and the theoretical seismograms is reached over the portion of the record containing the diffracted Sn and early reflected arrivals, see Figure 2.1.1.

4. Alignment between the horizontal component of source motion deduced from the direct S wave recorded at a single station (BRK) and the strike of the regional faulting was found. This phenomenon was observed only from sources within 100 km. Remarkably, the SV particle motion from these events is roughly linear. Assuming a homogeneous crust, plane wave theory would predict elliptical SV motion from shallow sources at these distances. Further research concerning the detailed crustal velocity structure and the theoretical phase relationship between SV and SH from a nearby point source is suggested.

5. The study indicates that in the routine location of regional ($100 \text{ km} < \Delta < 1000$ km) earthquakes using the S-P interval and short-period instruments located in west central California, it be assumed that the most obvious initial S arrival is either direct S or the first reflection from the crust-mantle boundary both traveling with an apparent surface velocity of about 3.6 km/sec.



Earthquake: June 26, 1968

Epicentral distance	309 km
Origin time	$01^h\ 42^m\ 19.5^s$
Azimuth	N46°W
Magnitude	5.5

Instrument:

Ultra-long period, horizontal.
See Figure A.3.

Theoretical Seismogram:

Epicentral distance	310 km
Source depth	5 km
Layer thickness	10 km

Constitutive parameters same as case of Figure 5.1

Figure 2.1.1

6. Care must be taken in attributing the lack of a strong SH component of Sn motion to attenuation or absorption during propagation, when the observations are taken from short-period instruments. The assumption of smooth source functions and plane, sharp boundaries gives rise to a long-period SH component of Sn motion. Observed high frequency content in Sn may be due to complexity in the crust-mantle boundary, a general conclusion reached by Červený and Helmberger for Pn, rather than solely due to a lack of absorption.

7. Suites of theoretical seismograms computed from the theory developed have shown the effect of variation of certain geometric and constitutive parameters. The main results are:

(1) An increase in the layer thickness when the source depth is fixed within the layer decreases the duration and increases the relative amplitude of Sn.

(2) An increase in source depth within the crust has little effect on the form of the seismogram.

(3) An increase in the layer velocity in the case of the crustal source decreases the duration and increases the relative amplitude of the diffracted, Sn, motion.

(4) Increasing the half space velocity in the case of the crustal source increases the duration and decreases the relative amplitude of Sn.

(5) Sources within the half space give rise to a pulse-like arrival after which the surface motion dies off rapidly.

All of the effects on the Sn motion are explainable physically through the consideration of Huygen's principle. For this reason, Sn is here considered a diffraction phenomenon. Certain empirical relations, taken from the theoretical seismograms computed for a layer thickness of 20 km, show a linear increase with epicentral distance in the number of cycles and period of the first half cycle of Sn motion in the range $300 < \Delta < 800$ km.

2.2 California Seismicity and Tectonics.

An exhaustive investigation using recent observations has been completed and the results published in the Bulletin of the Seismological Society of America.

Some 300 regional earthquakes have been located for 1962 through June 1965 using readings from local stations. During this period the telemetry network gave a uniformly high resolution for local earthquakes. The three greatest concentrations occur close to the north-facing Gorda escarpment off Cape Mendocino (latitude 40°.4 N); in the Coast Ranges west of the San Andreas fault (latitude 36°.8 N) between Watsonville and Salinas; and 40 km southwest, in the Gabilan Range west of the San Andreas fault near the Pinnacles (latitude 36°.5 N).

For 280 km northward of a small cluster which straddles the San Andreas off San Francisco, earthquakes were not detected near to the San Andreas but eastward in the Coast Ranges. The bulk of earthquakes with magnitudes in excess of 2-1/2 in the Coast Ranges have focal depths less than 5 km; no reliable depths were estimated greater than 20 km.

Seismicity on the Gorda scarp is high; it decreases westward along the Mendocino escarpment. Earthquakes were detected westward to the northerly-trending (Escanaba) trough which traverses the Gorda ridge and then northward along the Gorda ridge. Earthquakes also were located within the Gorda basin between the California coast and the Gorda ridge.

The first motions of P were read for 29 earthquakes using mainly the high-resolution stations of the central California array. The patterns are consistent with dominant right-lateral strike-slip along the San Andreas-Calaveras fault system from Parkfield to San Francisco Bay; some exceptional motion occurs near the San Andreas fault in the vicinity of the San Francisco Peninsula. In the Coast Ranges north of San Francisco Bay, right-lateral strike-slip appears again to be the rule.

Along the Gorda wall, the P polarities are consistent with the local escarpment trend of about N70°W. Earthquakes on the continental shelf north of Cape Mendocino and in the Gorda basin have fault-plane solutions which again indicate right-lateral strike-slip with strikes of about N40°W. The motions are consistent with a broad system of branching faults whose trend at the northern extremity becomes aligned with the Blanco fracture zone.

On the Gorda ridge, the shock of 18 April 1965 shows normal fault characteristics. Mechanisms for four earthquakes on the Blanco fracture zone are not uniform; right-lateral strike-slip motion is again found for the earthquakes of 8 May 1968 at 128° west and of 14 June 1965 at nearly 130° west, while an earthquake nearly mid-way between them has normal faulting.

Overall, from the Gulf of California to the Juan de Fuca ridge, seismicity and inferred earthquake mechanism agree well with a mechanical model based on the theory of J. Tuzo Wilson. The San Andreas fault decouples the continental United States from the western Pacific basin and permits essentially rigid-body motion of the adjacent regions. North of Cape Mendocino, a lack of parallelism between the Mendocino and Blanco fracture zones produces normal pressures across the Mendocino scarp which retards the local rate of displacement; as a result, some motion is transferred by transcurrent faulting from the San Andreas fault across the Gorda basin to the Blanco transform-fault system. The Gorda ridge is moving westward relative to the continental shelf. This ridge may with time become dormant from the south; further faulting with episodes of seismic activity may link the San Andreas fault near the Gorda escarpment with the Blanco fracture zone.

2.3 Frequency Spectral Analysis of Recordings from Array Stations to Estimate Crustal Properties.

R.A. Phinney has shown that the ratio of the vertical spectrum to the horizontal spectrum for teleseismic P-waves depends on the crustal structure beneath the recording station. Crustal structure may thus be inferred by a comparison of these experimental ratios with similar theoretical ratios for layered crustal models computed by means of the Haskell matrix theory.

While Phinney was able to estimate total crustal thickness from long-period data, attempts to determine detail within the crust from short-period data have met with only limited success. The lack of experimental ratios over a wide spectral range has been a major limiting factor in several of the crustal studies published. The magnetic tape data acquisition system, complemented by the analog-to-digital conversion unit recently installed, provides the Berkeley array with data over a spectral range of 0-5 cps.

The inability to isolate the effect of a portion of the crustal model in a band of the available spectral range has been a major technical difficulty. However, it seems plausible that the effect of a crustal layer will be periodic in the spectral ratio. By examining periodicities in the experimental ratios, it should be possible to isolate the effect of a section of the crust. Preliminary numerical calculations with theoretical ratios are encouraging. Because of resolution considerations, the full spectral range available with the Berkeley array is necessary to exploit this periodicity.

Examples of such experimental spectral ratios (see Figure 2.3.1) have been computed from the P-wave data of two South American earthquakes recorded on the BRK broad-band system.

Event	Station-to-Epicenter Azimuth	Distance	Magnitude (M_B)	Focal Depth
3 Nov. 1965	123°.4	66°.6	6.3 (BRK)	583 km
19 June 1968	125°.7	60°.2	6.7 (BRK)	28 km

The spectral ratios correlate well at the low frequencies and, although the correlation breaks down somewhat at the high-frequency end of the spectrum, it should improve considerably when the spectral effects of microseisms and other background noise have been removed.

The vertical spectral components (Figure 2.3.2) and the horizontal (radial) spectral components (Figure 2.3.3) that ~~have~~ have been used to form these experimental ratios clearly do not show the same degree of correlation. This effect can be attributed to the particular spectral character of the P-wave from each event. This input spectra, a function of the earthquake source mechanism and the ray path, is unknown and is removed by forming the spectral ratio. The dominant character of the input spectra can be clearly seen by comparing the vertical and horizontal spectra for the same event. The crustal effect on a component spectrum is then only a perturbation on the input spectrum, and can only be isolated by forming the ratio of the spectral components.

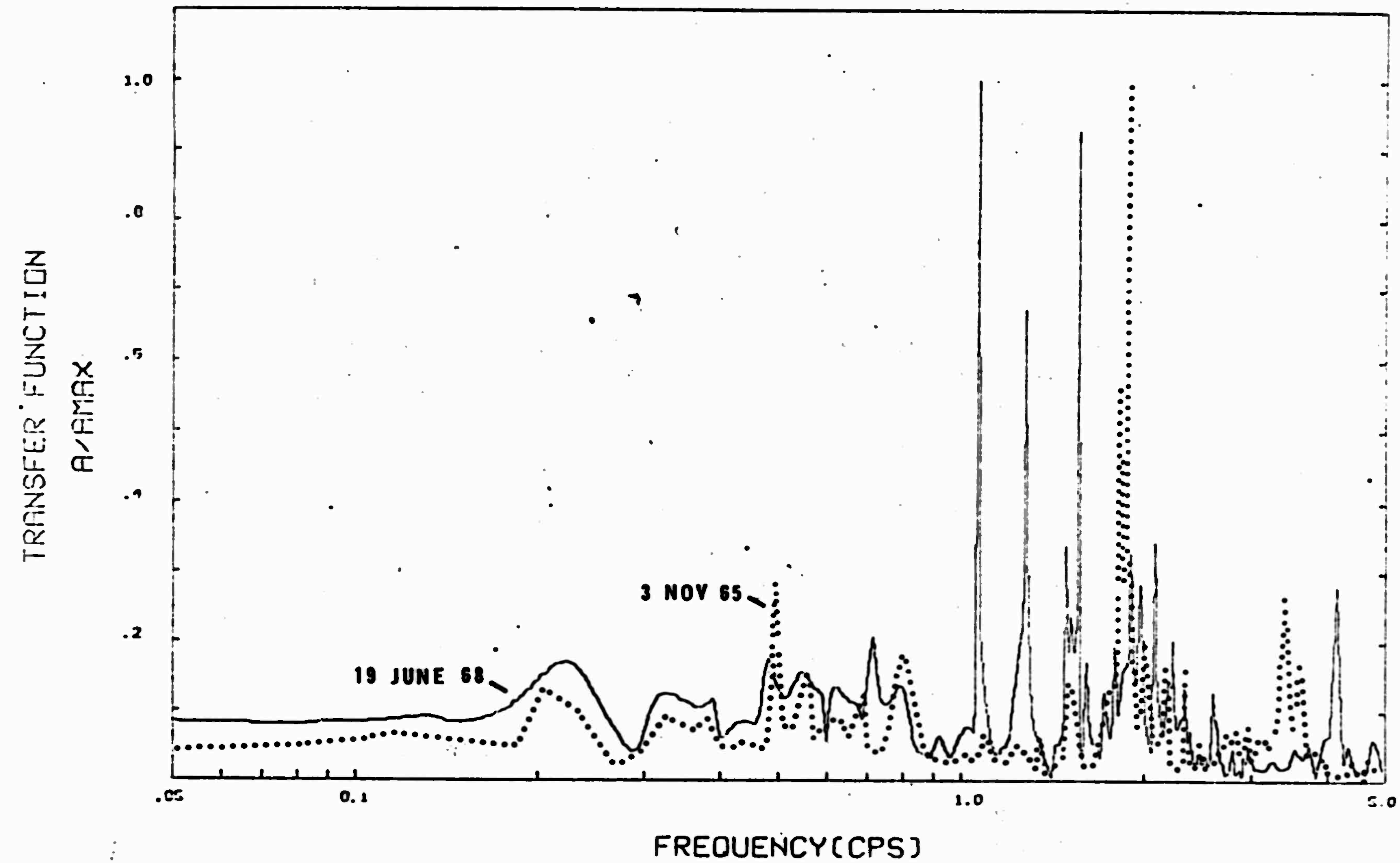


Figure 2.3.1. Empirical spectra of the ratio of the vertical component to the radial component of ground motion in the P wave train.

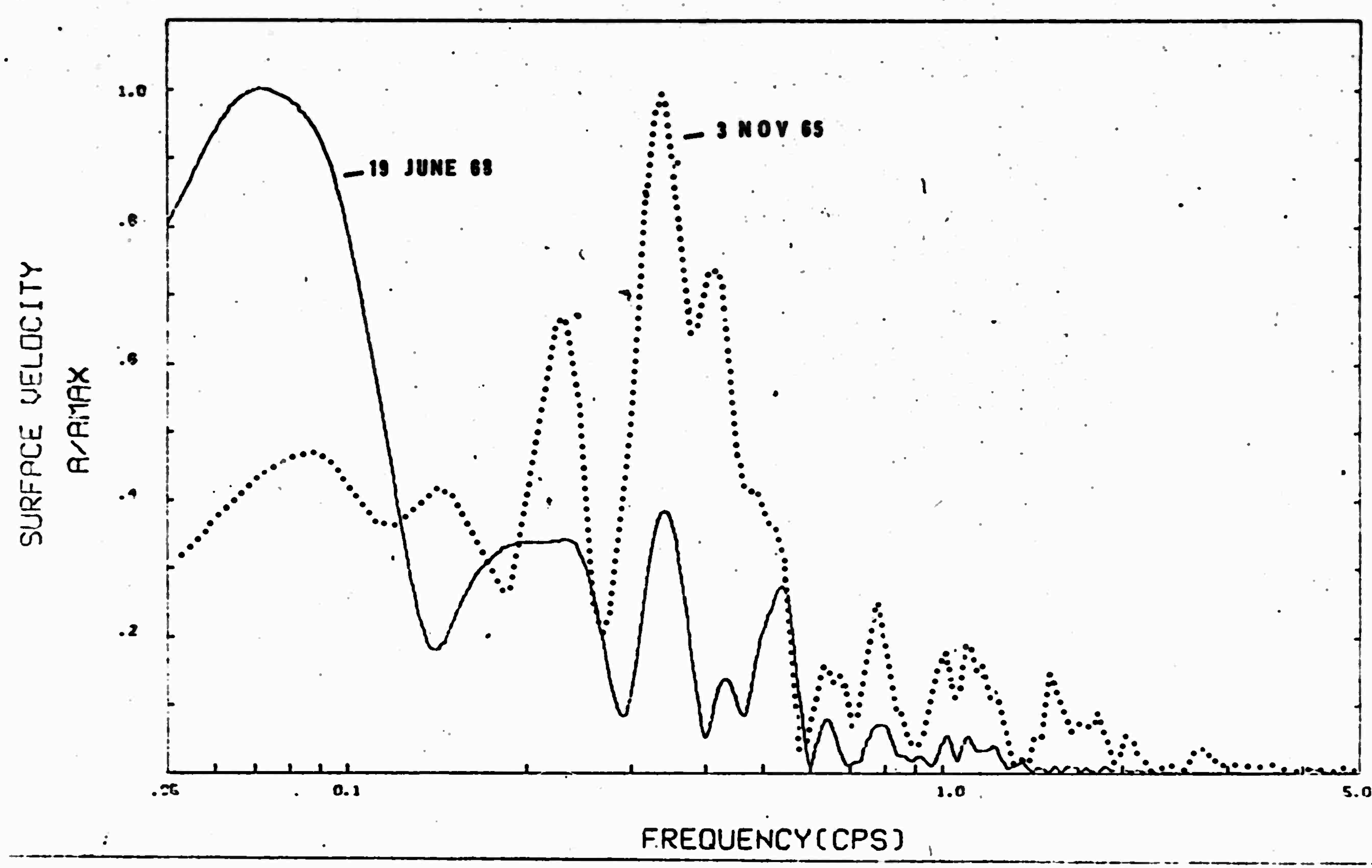


Figure 2.3.2. Fourier spectra of vertical components.

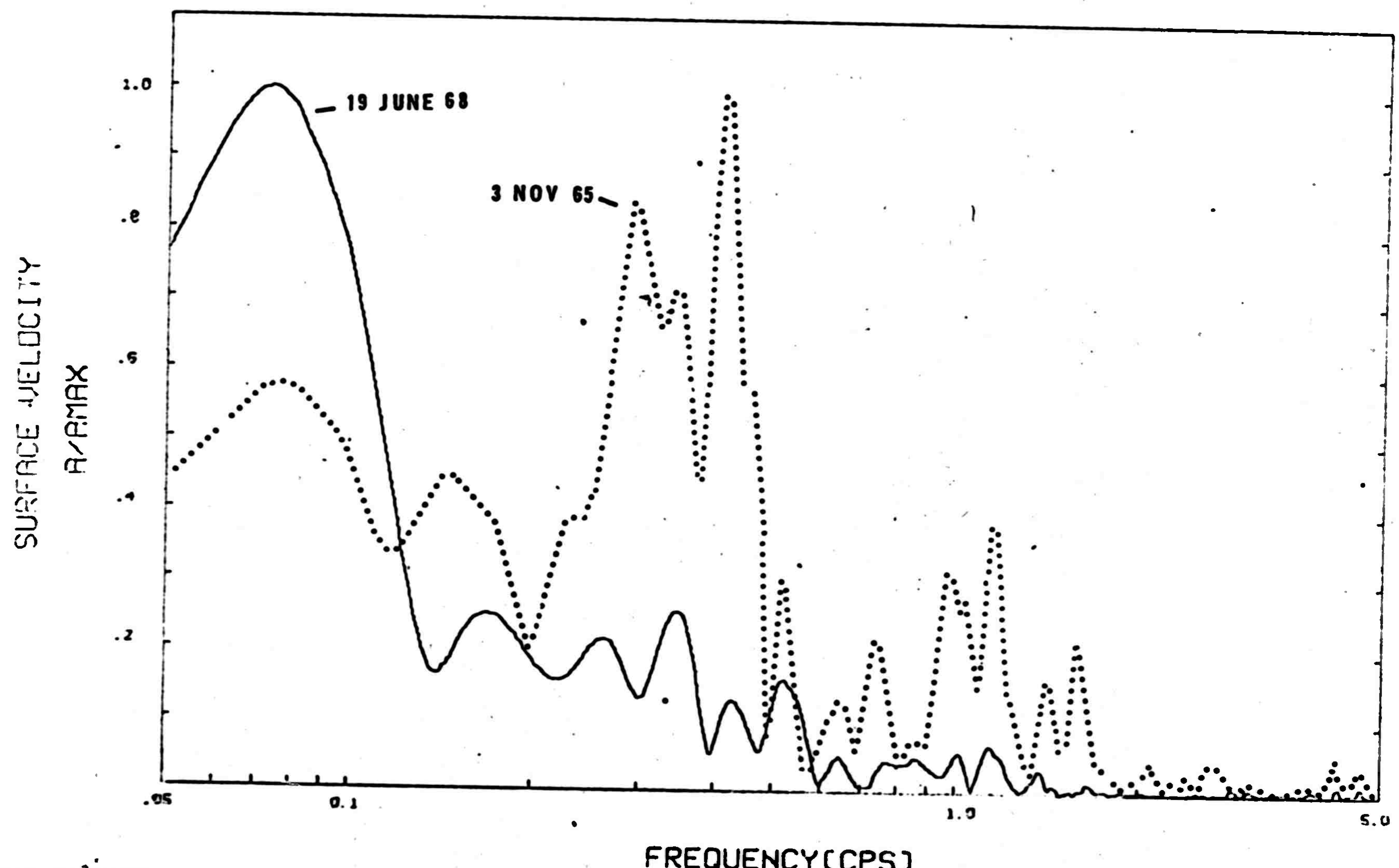


Figure 2.3.3. Fourier spectra of horizontal components.

The component spectra shown illustrate the spectral character of different earthquake sources. Since the ray paths for the two events are nearly the same, these spectral differences must reflect different focal mechanisms. For example, the dominance of the low-frequency end of the spectra for the 19 June 1968 event, in contrast with the spectra shown for the 3 November 1965 event, can be attributed either to its greater magnitude or to its shallower focus.

2.4 Detailed Crustal Structure within Array.

Several aftershock sequences within the Coast Ranges have been studied in detail using portable stations in the epicentral region. The hypocenters (with rather precise depths) for larger shocks in these sequences provided reliable Pn travel-time data that were lacking in the previous attempt (McEvilly, 1966) at a three-dimensional refraction interpretation for Coast Range crustal structure. In addition, more explosion data were generated by the U.S. Geological Survey program in the area. New data, earthquakes plus explosions, increased the number of source-station pairs available for a least-squares analysis from about 60 to 254. It was hoped that this increase would improve the definition of the lateral velocity change across the San Andreas fault zone and thus allow reinterpretation of azimuth and slowness anomalies across the array (Otsuka, 1966) in light of lateral change in crustal structure.

The technique assumes linearly increasing crustal velocity (with an optional additional constant velocity layer) over a constant velocity mantle with a variable depth (the Mohorovicic discontinuity is represented by a quadratic surface in latitude and longitude). A least-squares fit yields crust and Pn velocities and the configuration of the lower boundary to the crust. With sufficient data, separate analyses can be performed east and west of the San Andreas.

The best solution for the entire region, with an r.m.s. residual of 0.4 sec, is

Crust $V = 5.14 + 0.06\, Z$ km/sec

Z = depth in km

Mantle $V = 7.85$ km/sec

This contrasts with $5.16 + 0.08\, Z$ given by McEvilly in 1966.

The crustal thickness is close to 21 km at Berkeley with a rather uniform M discontinuity dip of 5° to 6° away from the coastline and a general crustal thickening to the southeast.

Crustal arrivals indicate the velocity distribution in the upper 5 to 10 km of the crust (roughly the depth range of the earthquakes used) varies across the San Andreas zone as follows:

West	East
$V = 5.33 + 0.03\, Z$	$V = 4.85 + 0.1\, Z$

At a depth of about 7 km the difference is gone and the average model listed above applies. While the travel-time difference for vertical incidence is less than 0.1 sec in these two models, essentially horizontal propagation (Pg) from shallow events can involve differences exceeding 0.5 sec.

The average M discontinuity configuration is influenced by these differences across the San Andreas zone. The effect is roughly an apparent increase of Mohorovicic dip away from the coast of 1° to 2°, meaning the actual average dip is more like 4° or 5° inland.

Implications of this structural model on an interpretation of azimuthal variations in slowness across the array can be estimated. For the geometry of the Berkeley array the model suggests that some ± 0.2 sec per degree variation in apparent slowness of teleseismic P waves can be expected from crustal structure alone. Otsuka (1966) found some ± 0.7 sec per degree for the array. Thus it would appear that a significant part of the anomaly is due to lateral variations in the upper mantle.

References

McEvilly, T.V. (1966). Crustal structure estimation within a large-scale array, Geophys. Journal, 11, 13.

Otsuka, M. (1966). Azimuth and slowness anomalies of seismic waves measured in the central California seismographic array, Part I, Observations, Bull. Seism. Soc. Am., 56, 223.

2.5 Crustal Profile from BOXCAR and BENHAM across Great Valley and Coast Ranges.

During 1968 a joint program was begun by the Berkeley Seismographic Station and ESSA's Earthquake Mechanism Laboratory to record and interpret travel times from large-yield nuclear tests in Nevada. The first test to be so utilized was the NTS BOXCAR occurring at 1500 GCT on April 26, 1968. A series of eight temporary stations was established which, together with seven permanent stations, provided a profile passing through Berkeley which extended from Yosemite Valley (station YOS) in the Sierra Nevada to Lamont's Ocean Bottom Seismometer (station OBS) about 135 km southwest of Point Arena (see Figure 2.5.1).

Due to the high explosive yield, generally excellent Pn arrivals were recorded at all stations together with P waves which travelled only through the crust (Pg-type). The quality of the latter seemed to vary with crustal structure. A plot of the measured Pn travel times is given in Figure 2.5.1. Table 1 lists these times together with coordinates of the stations. As described in the Quarterly Report for April-June 1968, an attempt to interpret the data as due to a crustal model consisting of a flat layer over a half space showed that there were anomalous time delays of up to 1.4 seconds at stations in the Sacramento Valley (380 $< \Delta <$ 450 km), i.e. PDM, RIV, EUN, YAM. Eliminating these stations from the data, linear least-squares equations fitted to the remaining data points (corrected for elevations of shot point and seismometers) for both Pn and Pg gave the following relations:

$$\text{Pg} \quad t = \Delta/5.99$$

$$\text{Pn} \quad t = \Delta/7.87 + 6.2.$$

For the assumed model this implies a Pg velocity of 5.99 km/sec, a crustal thickness of 28.7 km, and velocity at the top of the mantle of 7.87 km/sec.

The large time delays for the Sacramento Valley could then be interpreted in several ways assuming the valley to be a perturbation of the above model. Eaton's (1963) model of valley sediments as a layer of up to 4.5 km of velocity $v = 1.9 + 0.55z$ (z depth in km) over a crust of velocity 6.0 km/sec does not predict sufficiently great time delays unless the Mohorovicic discontinuity is allowed to warp downward by an amount equal to the thickness of the sedimentary layer. Alternately, the data could be explained by sediments of constant velocity (3.5 km/sec) over a 6 km/sec crust with a flat horizontal base if the sediments exceeded 9 km in depth.

A surprising result from the BOXCAR data was the apparent lack of an obvious mountain root; this appeared in the form of small time delays at stations in the Sierra Nevada, YOS and ELP. In order to recheck the travel times for stations in this region two temporary stations, OCT and HAP, were established for the nuclear test BENHAM which occurred at 1630 GCT on December 19, 1968. In addition, a station was established at Mammoth to the east of the Sierra Nevada crest by Dr. Dean S. Carder of ESSA's Earthquake

Mechanism Laboratory. The data from OCT and HAP confirmed the previous data from BOXCAR but the arrival at Mammoth (MAM) seemed to show a significant delay of 0.7 sec, an indication, perhaps, of the Sierra root (Figure 2.5.1). Even if the elevation difference between Mammoth (8900 ft.) and Yosemite (4040 ft.) is taken into account the delay at Mammoth with respect to Yosemite is still more than 0.5 sec.

The detailed analysis of the data is being carried out in collaboration with Dr. Carder and plans are being made to extend the profile over the crest of the Sierra Nevada during future nuclear tests in order to better define the crustal configuration under these mountains.

Table 1

STATION COORDINATES AND Pn TIMES

BOXCAR

Station	Lat.	Long.	Δ(km)	Elev. (Meters)	Raw Travel Time (T) m	s	T-Δ/7.87-6.2
YOS	37° 45.1'	119° 35.3'	282	1230		42.1	+0.1
ELP	37° 40.5'	119° 46.9'	297	610		44.1	+0.1
COU	37° 42.7'	120° 11.9'	334	534		48.7	+0.1
JAS	37° 56.8'	120° 26.3'	359	457		52.2	+0.3
*PDM	37° 46.9'	120° 46.5'	386			56.2	+1.0
*RIV	37° 48.2'	121° 00.9'	407			58.9	+1.0
*EUN	37° 49.3'	121° 14.0'	426		1	01.8	+1.45
*YAM	37° 50.4'	121° 26.9'	445		1	04.0	+1.2
*GRU	37° 51.5'	121° 41.8'	467		1	05.8	+0.25
*MDC	37° 52.8'	121° 55.1'	487		1	08.2	+0.15
BKS	37° 52.6'	122° 14.1'	515	276	1	11.6	+0.0
BRK	37° 52.4'	122° 15.6'	517	81	1	11.8	-0.1
*SFR	37° 47.3'	122° 23.4'	527		1	13.2	-0.0
*OLC	38° 02.4'	122° 47.5'	565		1	17.9	-0.1
**OBS	38° 09.2'	124° 54.4'	751	-3830	1	41.3	-0.3

BENHAM

Station	Lat.	Long.	Δ(km)	Elev. (Meters)	Raw Travel Time (T) m	s	T-Δ/7.87-6.2
*MAM	37° 39.1'	119° 02.2'	232	2710		36.3	+0.7
HAP	37° 43.8'	119° 33.5'	278	1230		41.6	+0.0
OCT	37° 35.5'	119° 57.7'	311	503		46.0	+0.25
JAS	37° 56.8'	120° 26.3'	359	457		52.0	+0.2
BKS	37° 52.6'	122° 14.1'	514	276		11.4	-0.1

*ESSA Stations.

**Ocean Bottom Seismometer - Lamont Geological Observatory.

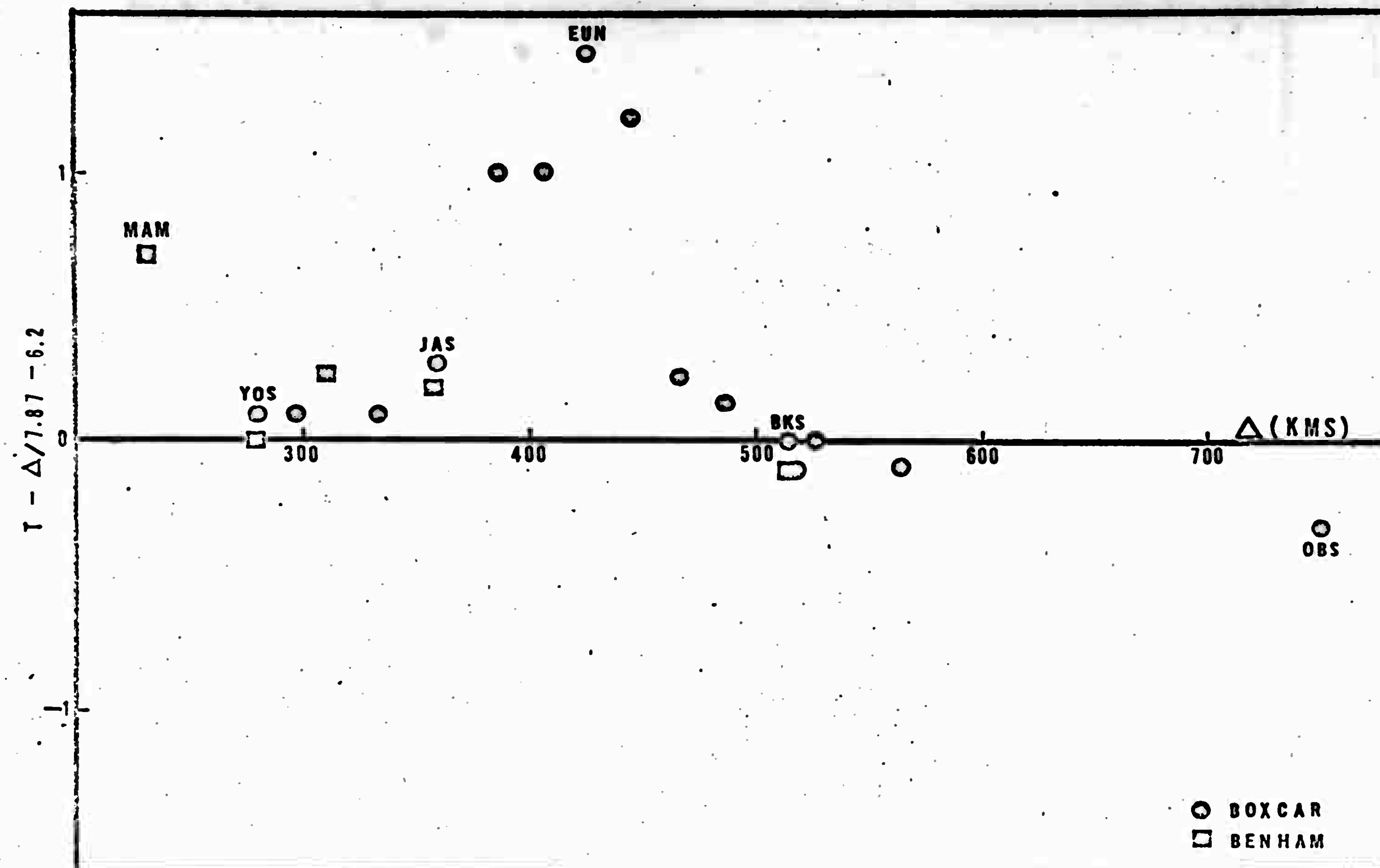


Figure 2.5.1. Reduced Travel-Times from BOXCAR and BENHAM.

2.6 Use of the Array for Studies of the Earth's Interior.

Longitudinal waves have been analyzed which are observed at distances between 105° and 115° some minutes after the arrival of the P waves diffracted by the mantle-core boundary. A number of seismograms from the Berkeley network which clearly show these wave trains have been examined.

The observations suggest (i) that discernible PKiKP waves with wavelengths of order 10 km are reflected from the boundary of the Earth's inner core back to distances of at least 105°, and (ii) that many longitudinal wave onsets (the PdP phase) having travel-times up to 60 sec before PKiKP and 90 sec before PP arrive near 110° by means of reflection from the lower side of physical discontinuities in the upper mantle of the Earth.

The first result is consistent with a relatively sharp increase in P velocity between the transition zone and the inner core at a radius of about 1220 km. The second suggests the existence of a number of discrete shells of different elastic properties in the Earth's upper mantle; in particular, one prominent group of PdP waves of order 2 sec period is consistent with reflection from a discontinuity near 400 km. The result provides confirmation of the overall high velocity gradient near this depth inferred by L. Johnson; there is an indication, however, of first-order discontinuities in the velocity function assumed to be smooth by Johnson.

In related work, a set of travel times for the core-waves PKP was derived from observations and tabulated. The Berkeley array of stations provided necessary important readings in this study. All values are tied to the 1968 P travel times for a surface-focus for $75° < \Delta < 95°$ due to

Herrin et al. Times for both the branches DF (110° <Δ< 180°) and GH (125° <Δ< 156°) are listed in more detail than previously, together with an extension for PKiKP waves, from 110° back to 105°. Checks on the DF branch for 120° <Δ< 150° using observations from the 1965 LONGSHOT explosion provide no evidence for further revision of the first section of this branch.

Special attention is given to the times of the DF branch (PKIKP) for Δ > 150°. Delays at 153° and 162° of order 3 sec reported by K. Ergin in 1967 are not confirmed. There is a group of small arrivals about 1 sec earlier than predicted by the present PKIKP tables for 156° <Δ< 160°, approximately. These may indicate an error in the curvature of the present curve and, perhaps, a slight discontinuity in the inner core near a radius of 850 km. No evidence is found for a layer of any consequence with reduced P velocity in the inner core.

Revised times for the branch AB (denoted by PKP_2) are based mainly on readings from the Indian Ocean earthquake of February 17, 1966 and the Jeffreys-Bullen times. For Δ> 155°, the new empirical times confirm the curvature of the Jeffreys-Bullen curve for PKP_2; however, near Δ = 152°, the observed curvature appears to be greater than that of the latter curve. The new times are of the order of one second earlier; this result is highly sensitive to the precision of the epicenter estimated for the 1966 shock. PKP_2 is traceable at least out to 185°.

Depth allowances for PKP have been computed, by T. Qamar, to provide a set consistent with the model Earth used by Herrin et al in the 1968 P tables.

Very precise location of hypocenters remains a leading seismological objective. Core phases have two characteristics which provide powerful constraints on location. The first, which is often noted, arises from the general sharpness of the PKIKP onset as a first arrival in certain distance ranges, particularly for $\Delta > 160°$. At these distances, $dt/d\Delta$ is so small that the arrival time is almost entirely a function of depth and origin-time. The second, which appears not generally known, is a consequence of the great contrast in slope between the AB and EF branches for $\Delta > 160°$. Near 160°, the ratio exceeds four to one, or twice that of the slopes of S and P near 50°. An error of 0.1° or about 10 km in epicentral distance amounts to a residual in PKP_2 - PKIKP of order 0.3 to 0.5 sec which is now detectable. Unfortunately, no underground nuclear explosion with known elements has yet occurred at a site which has a number of stations at $\Delta > 150°$. Large shots at the Nevada Test Site have a few antipodean seismographic stations in South Africa, Malagasy and Kerguelen. This calibration would be most valuable.

2.7 Post BOXCAR and BENHAM Seismic Events.

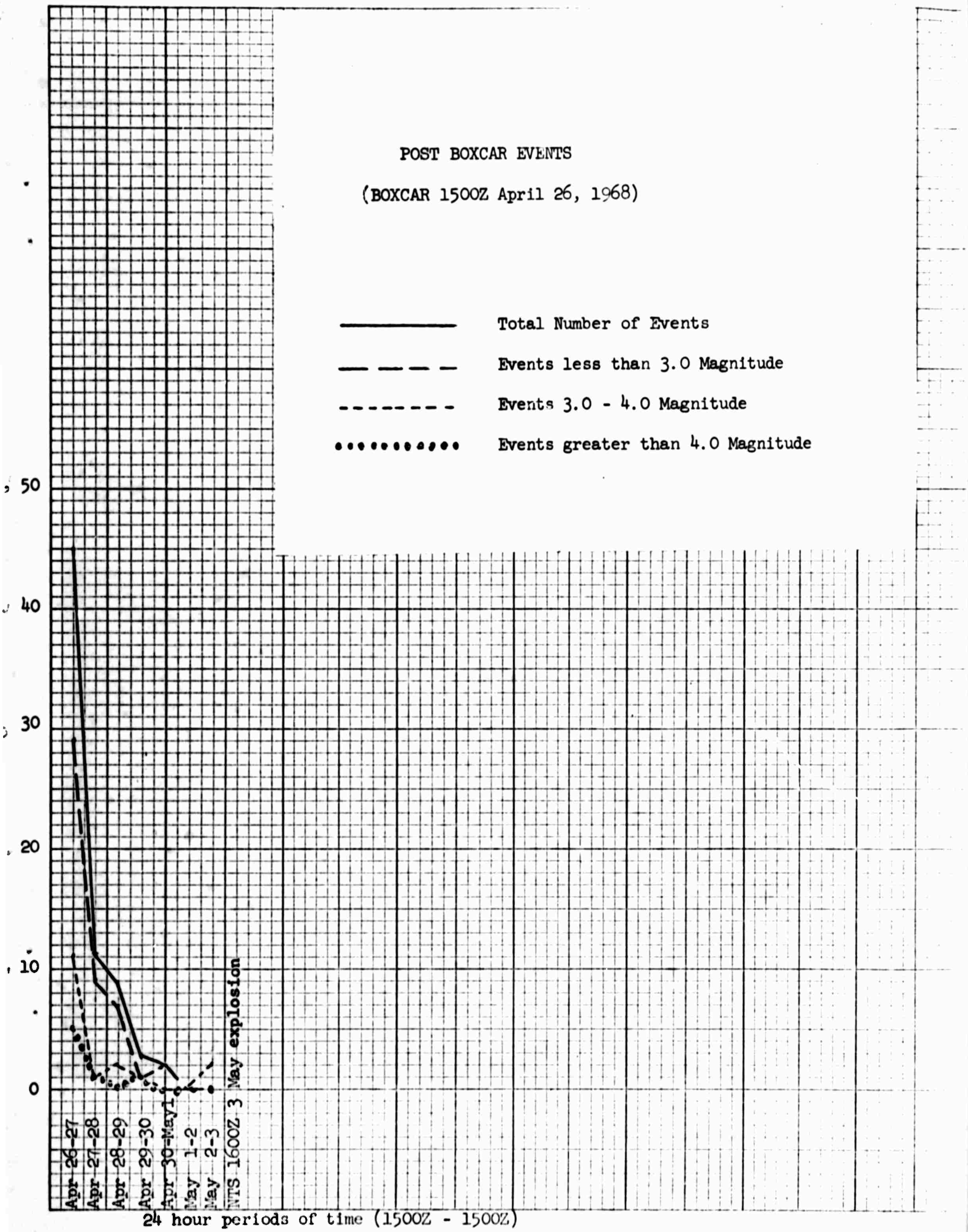
During 1968 two large underground nuclear explosions were detonated at the Nevada Test Site, BOXCAR on 26 April at $15^h00^m00.1^s$, and BENHAM on 19 December at $16^h30^m00^s$ GMT. The coordinates of BOXCAR were 37°.295 N, 116°.456 W and of BENHAM 37°.232 N, 116°.474 W; the sites were thus some 7 km apart, BOXCAR to BENHAM azimuth is 193°. Both shots, which had magnitudes of 6.4 and 6.3, respectively, as determined by Wood-Anderson seismographs of the Berkeley network, were followed by a large number of seismic events in the source region, much like the aftershock sequences associated with natural earthquakes. Something similar had been observed in the past for large shots such as GRE EY and FAULTLESS; the Berkeley network regularly records "collapses" following underground explosions in Nevada. The Jamestown station is sensitive to these events down to equivalent magnitudes less than 2.0. Seismic sources of magnitude 3.0 and above are sufficiently well recorded over the Berkeley array that the location can be placed with confidence in the test site.

Because of the high interest in this phenomena of aftershocks from explosions, it seemed worthwhile to tabulate the statistics of occurrence of such events. Figures 2.7.1 and 2.7.2 present the data for BOXCAR and BENHAM with classification as to magnitude range. The late burst of activity for BENHAM is apparently associated with fracturing a few km south of the explosion.

In an attempt to find characteristics of the BENHAM aftershocks distinguishing them from the main shot, the phasesof the 10- to 30-seconds period Rayleigh waves as recorded on the Berkeley broadband systems were compared. Two interesting features were found:

1. Initial phase of Rayleigh waves from aftershocks is identical to that of BENHAM. This is in marked contrast with the $\pi/2$ to π phase difference seen in normal collapse events (e.g. HALFBEAK, DUMONT).
2. Surface wave to body wave energy ratios for aftershocks ($M > 4$) in the first week are essentially identical to the ratio for BENHAM. However, for the cluster of events January 8-11 several km south of the shot site, the proportion of energy in the Rayleigh wave relative to body waves is nearly an order of magnitude lower than for the earlier group.

These observations clearly hold information on the mechanics of strain relief by after events following large nuclear explosions. The low level surface wave radiation in the aftershocks of BENHAM has important nuclear detection implications. The late aftershocks certainly do not appear, in a Ms/Mb sense, to be earthquakes. Further work along these lines will be carried out in the new contract period.


POST BOXCAR EVENTS
(BOXCAR 1500Z April 26, 1968)
Total Number of Events
Events less than 3.0 Magnitude
Events 3.0 - 4.0 Magnitude
Events greater than 4.0 Magnitude
50
40
30
20
10
0
Apr 26-27
Apr 27-28
Apr 28-29
Apr 29-30
Apr 30-May 1
May 1-2
May 2-3
NTS 1600Z 3 May explosion
24 hour periods of time (1500Z - 1500Z)

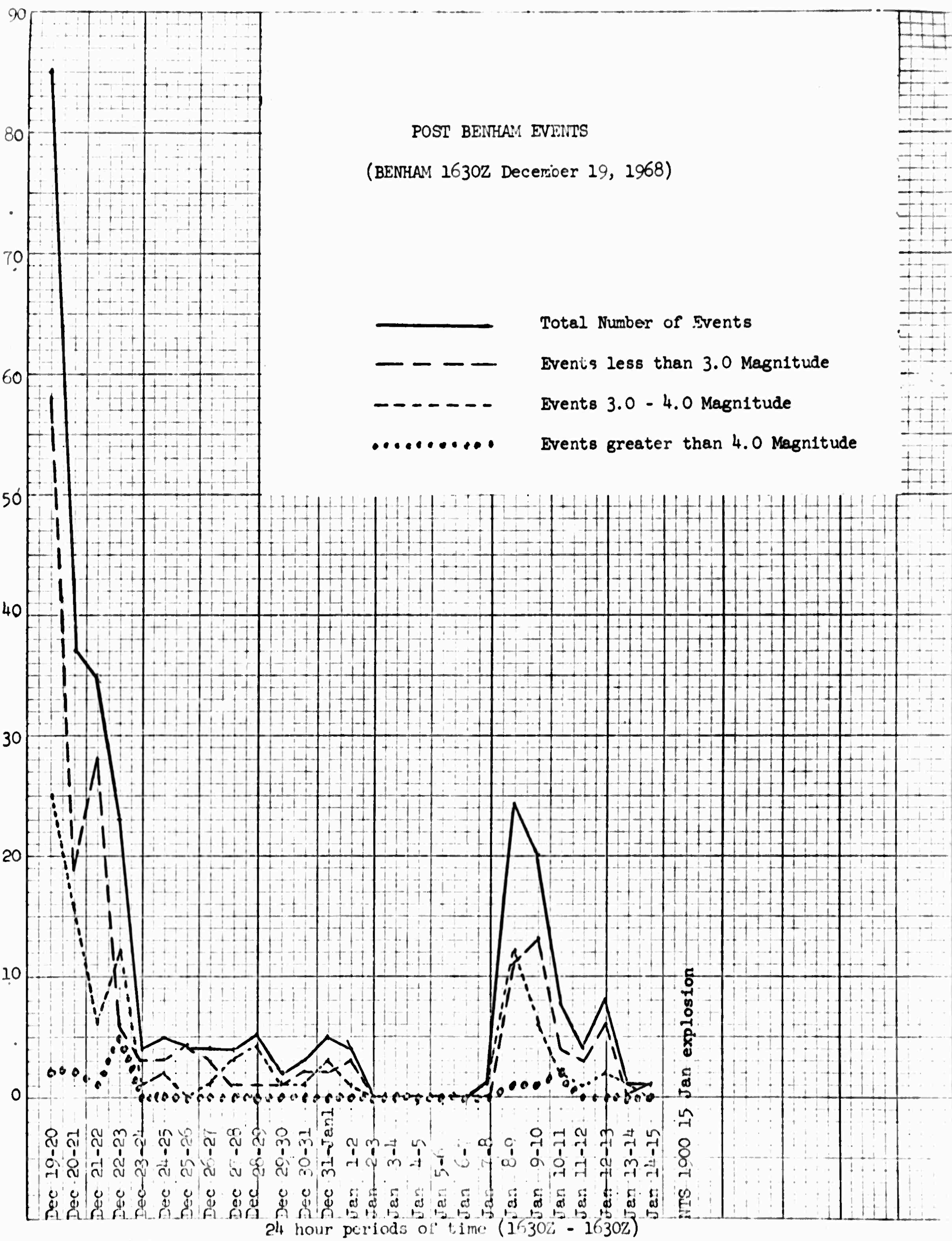

POST BENHAM EVENTS
(BENHAM 1630Z December 19, 1968)
Total Number of Events
Events less than 3.0 Magnitude
Events 3.0 - 4.0 Magnitude
Events greater than 4.0 Magnitude
0
10
20
30
40
50
60
70
80
90
Dec 19-20
Dec 20-21
Dec 21-22
Dec 22-23
Dec 23-24
Dec 24-25
Dec 25-26
Dec 26-27
Dec 27-28
Dec 28-29
Dec 29-30
Dec 30-31
Dec 31-Jan1
Jan 1-2
Jan 2-3
Jan 3-4
Jan 4-5
Jan 5-6
Jan 7-8
Jan 8-9
Jan 9-10
Jan 10-11
Jan 11-12
Jan 12-13
Jan 13-14
Jan 14-15
NTS 1900 15 Jan explosion
24 hour periods of time (1630Z - 1630Z)

Figure 2.7.2.

3. The Network Operation.

Two stations in northern California were added to the Berkeley telemeter network (with non-AFOSR funds) during the period covered by this report. No other changes were made to the system. It has operated satisfactorily, with down time of less than one per cent. All records have been processed, catalogued, and are in library storage.

One of the new telemetered stations is located at the Mineral photographic station in Lassen Volcanic National Park (station designation MIN). The station began recording on trace 10 of the develocorder on October 14, 1968. The other new station, Fickle Hill, California (FHC), located on siltstone overlying greywacke, is nine miles southeast of our photographic station at Humboldt State College in Arcata, California. Coordinates of this station are 40° 48' 8.5" N, 123° 59' 3.3" W. The signal from the Benioff vertical component seismograph at FHC began recording on trace 15 of the develocorder on December 10, 1968; its magnification at 1 cps is about 60,000.

Table 1 lists the Berkeley telemeter array stations as they appear on the develocorder as of December 31, 1968.

Table 1

1.	Time		6.	JAS	Z	11.	MHC	Z
2.	PCC	Z	7.	SAO	N45°E	12.	GCC	N45°E
3.	SAO	Z	8.	BRK		13.	PRS	N45°E
4.	PRI	Z	9.	SAO	HFZ	14.	LLA	Z
5.	PSM	Z	10.	MIN	Z Oct. 14, 1968	15.	FHC	Z Dec. 10, 1968

The mobile (truck mounted) station was used for site testing in northern California. Portable stations were used to record the BOXCAR and BENHAM nuclear tests. The mobile station, which was the last of three assembled in 1962 from AFOSR contract funds, was dismantled on November 5, 1968. The instruments from this station were assembled into portable units that can be placed in a station wagon and taken to any site, thereby saving the truck rental. Figures 3.1 and 3.2 are calibration curves for these two portable units.

The magnetic tape data acquisition system (Minneapolis-Honeywell LAR 7400 analogue-slow speed recorders) has operated continuously since June 19, 1964, with 183 hours down time due to malfunctioning out of 37,408 hours of operation; this amounts to less than one-half per cent. In addition, down time of four hours bi-weekly is necessary for tape changing and dubbing of events on library tapes.

Table 2 lists stations and channel numbers of the magnetic tape system for the period January 1, 1968 to December 31, 1968.

Table 2

1. Time code	6. GCC N45°E	11. MHC N45°E
2. BRK ZSM	7. Comp	12. SAO N45°E
3. BRK Z	8. BRK N45°ESM	13. PRS N45°E
4. BRK N45°W SM	9. BRK N45°E	14. JAS Z
5. BRK N45°W	10. PRI Z	

The data acquisition system at Berkeley has been extended by the addition of continuous magnetic tape recording at the San Andreas Observatory (Geotech 21022 using 14-channel tape compatible with the LAR 7400 playback machine). More recently, an analogue-to-digital converter built by Datatron has been brought into service. Its maximum sampling rate is 1000 samples per second; each sample is a 12-bit binary word recorded on a Calma magnetic tape recorder.

Magnification curve for light weight portable set # 1
Helicorder Amplifier set at 0 db
One gram weight lift causes 31.5 mm deflection at 30 db

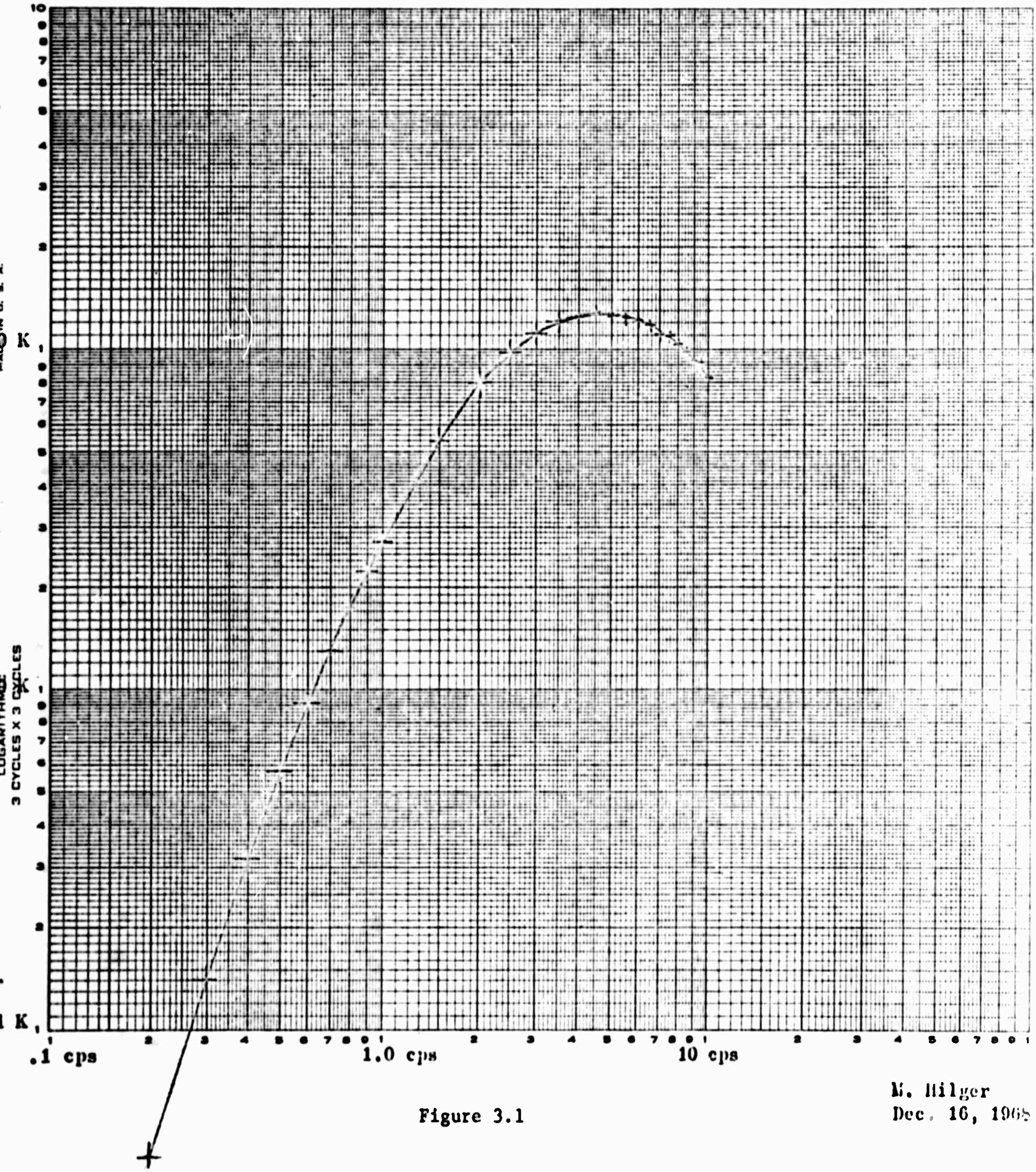


Figure 3.1

M. Hilger
Dec. 16, 1968

Magnification curve for portable set # 2
Helicorder Amplifier set at 24 db
One gram weight lift causes 27.5 mm deflection at 30 db

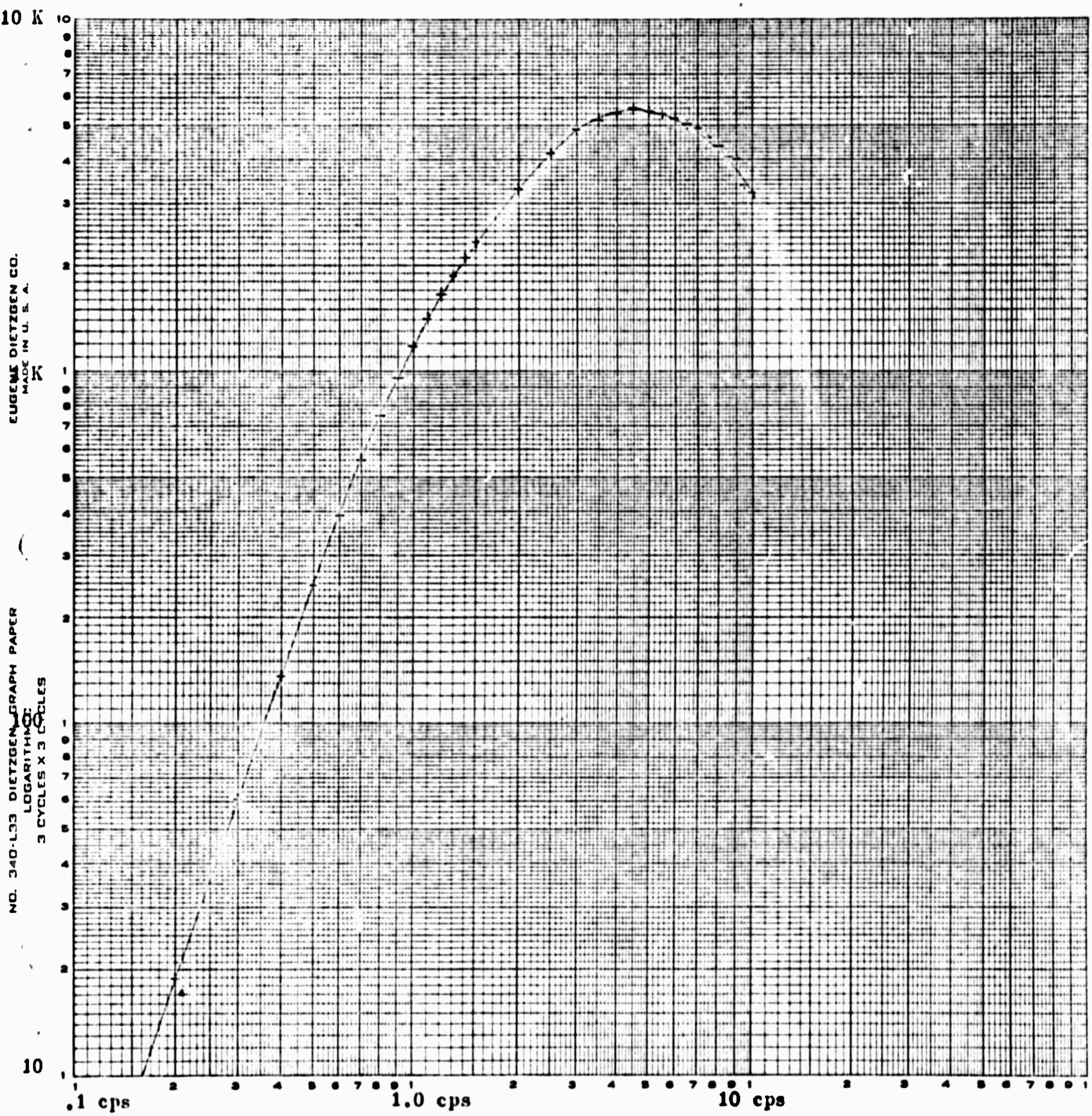


Figure 3.2

M. Hilger
Dec. 16, 68

4. Research Papers Depending Upon Contract Support.

The list which follows contains those papers which were presented at meetings, were published in 1968, or are in the press and which depended wholly or partly on observations from the UC, Berkeley array.

Bolt, B.A. (1968). "Estimation of PKP Travel-Times", Bulletin of the Seismological Society of America, 58, 1305-1324.

Bolt, B.A., C. Lomnitz and T.V. McEvilly (1968). "Seismological Evidence on the Tectonics of Central and Northern California and the Mendocino Escarpment", Bulletin of the Seismological Society of America, 58, 1725-1767.

Bolt, B.A., T.V. McEvilly and J.R. Filson (1968). "The San Andreas Geophysical Observatory: Initial Results", Meeting of American Geophysical Union, Washington, D.C.

Bolt, B.A., C. Lomnitz and T.V. McEvilly (1968). Earthquakes of the San Andreas Fault, Mendocino and Gorda Ridges Complex", Meeting of American Geophysical Union, Washington, D.C.

Bolt, B.A. and T.V. McEvilly (1968). "Earth Structure and Focal Mechanism: Evidence from the Berkeley Array", Comptes Rendus, Bur. Cent. Seis. (in the press).

Bolt, B.A., M. O'Neill and A. Qamar (1968). Seismic Waves Near 110°: Is Structure in Core or Upper Mantle Responsible?", Geophysical Journal of the Royal Astronomical Society, 16, 475-487.

Chuaqui, Luz and T.V. McEvilly (1968). "Crustal Structure within the Berkeley Array", Meeting of Seismological Society of America, Tucson, Arizona.

Filson, J.R. (1968). "SH Motion at Near Distances", Meeting of American Geophysical Union, Washington, D.C.

McEvilly, T.V. (1968). "Seafloor Mechanics North of Cape Mendocino, California", Nature, 220 (5170), 901-903.

Zanetti, J. and T.V. McEvilly (1968). "Characteristics of Microearthquakes Recorded at the San Andreas Geophysical Observatory", Meeting of Seismological Society of America, Tucson, Arizona.

5. Expenditures to Date (December 31, 1968).

General Assistance	$31,411
Supplies and Expense	38,102
Equipment and Facilities	4,224
Total	$73,737

Remaining funds are committed for January and February 1969 expenses involved in completing the two-year contract.