Microearthquake Studies Using Sonobuoys: Preliminary Results from the Gulf of California

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Summary

We report preliminary results of the use of telemetering sonobuoy hydrophones to record small local earthquakes in the Gulf of California. Standard naval sonobuoys were deployed, and their FM radio transmissions monitored and recorded on board ship. The airgun on the moving ship was used to triangulate to obtain the positions of the sonobuoys as they drifted.

In the six weeks of operation (only two of which were given priority for sonobuoy operation), we recorded numerous individual earthquakes, several earthquake swarms of 5-50 events and one swarm in the Guaymas Basin consisting of about 1000 events in a period of 6 hr. During the Guaymas Basin swarm an array of sonobuoys was deployed in the epicentral region shortly after the beginning of the swarm, and source locations have been determined to within 2 km. With further refinements of navigation and bathymetry we expect the source location error to be reduced further. Such accuracy has never before been achieved for this type of sequence and compares favourably with the accuracy of aftershock locations on land. The events in the Guaymas swarm were located in a 2 km wide graben associated with the Guaymas Basin spreading centre.

1. Introduction

Successful use of telemetering sonobuoy hydrophones to record small local earthquakes in the Gulf of California has demonstrated the feasibility of doing extensive, inexpensive seismicity studies in the oceans. During ‘Hypogene’ Expedition of R/V Melville (March and April 1972) we recorded numerous local earthquakes using this technique. As a result of these studies we conclude that it is possible to successfully record nearby earthquakes down to about magnitude 0. The character of the seismograms from the hydrophones is quite similar to the character of seismograms of small local earthquakes on land. Initial arrivals are sharp and the earthquake signal is easily distinguishable from noise. Thus the techniques of studying local seismicity on land can be applied to the oceans.

The studies of the type carried out here may be thought of as the ocean analogy of land microearthquake studies using portable seismographs. The telemetering sonobuoys are easily deployed and can operate at high sensitivity. The major difficulties not common with land studies are the necessity of remote recording, the drifting of the sonobuoys and the necessity of expensive deployment and recording.
vehicles (aircraft or boats). but these difficulties are not great when considered in
terms of the value of the scientific results which can be obtained.

In this paper we report the technique of sonobuoy recording used on 'Hypogene'
Expedition, together with our preliminary analysis of the results.

2. Pressure effects of undersea earthquakes, and the feasibility of hydrophone recording
of small earthquakes

In this section we discuss the pressure variations at nearsurface hydrophones
produced by seismic waves from an undersea earthquake, and relate this to the
magnitude of the earthquake.

First, we consider the pressure effect of a plane wave of given displacement amplitu-
de in the water. We consider only one frequency component. The relevant
equations are

\[ p = -KV \cdot s, \]
\[ s = a k \exp \left[ i(k \cdot r - \omega t) \right] \]

with

\[ K = \rho \alpha^2 \]

where \( a \) = displacement amplitude, \( K \) = bulk modulus, \( \rho \) = density, \( \alpha \) = velocity of
wave propagation, \( k \) = wave vector; \( k = |k| = \omega/\alpha \); \( \hat{k} \) = unit vector parallel to \( k \) so
that

\[ |p| = \rho \alpha^2 k a \]

\[ = \rho \alpha \omega \alpha. \]

Ignoring phase factors, the plane wave defined above is equivalent to a pressure
wave of amplitude \( p_0 = \rho \alpha \omega \alpha. \). The pressure amplitude produced by a seismic wave
in the sea floor of amplitude \( a \) is therefore

\[ P_0 = \rho \alpha \omega T a \]

where \( T \) is the amplitude transmission coefficient from rock to water corresponding
to the nature and angle of incidence of the seismic wave.

For a \( P \)-wave at normal incidence, \( T \) is given by the well-known simple formula,

\[ T = \frac{2 \rho_r \alpha_r}{\rho_r \alpha_r + \rho_w \alpha_w}, \]

where the subscripts \( r \) and \( w \) refer to rock and water respectively.

Taking typical approximate values

\[ \rho_w = 1 \]
\[ \rho_r = 2.5 \]
\[ \alpha_w = 1.5 \]
\[ \alpha_r = 6.0 \]

we have

\[ T = 1.8. \]

This implies that the interface motion is reduced only about 10 per cent from that of
a free surface (\( T = 2 \)), so that for magnitude calculations it is a reasonable approxi-
mation to ignore the effect of the water layer on the ground motion. Although this
has only been shown only for normally incident \( P \)-waves, the curves given in Ewing,
Jardetsky & Press (1957) for transmission and reflection at a fluid-solid interface
show that the approximation is in fact valid over a wide range of angles of incidence,
both for \( P \) and \( S \)-\( V \) waves.
We therefore assume that, for the purpose of magnitude calculations, the ground motion is unaffected by the presence of the water layer. Consider a plane seismic wave travelling in the $x$-$z$ plane. This wave causes a displacement at the boundary which is given by

$$u = u_0 \exp \left[ i \omega (x/c_1 - t) \right]$$

where

$$c_1 = \frac{\alpha_r}{\sin \phi} \quad \text{for a P-wave}$$

or

$$c_1 = \frac{\beta_r}{\sin \phi} \quad \text{for an S-wave.}$$

($\phi = \text{angle of incidence of impinging wave}$)

$u_0$ is not the amplitude of the plane wave, but of the surface motion produced by the wave and its reflections, and is the amplitude which would be recorded by an ocean-floor seismometer.

The propagating disturbance gives rise to a P-wave in the water, which is given by

$$s = k a \exp \left[ i \omega \left( \frac{z \cos \theta}{\alpha_w} + \frac{x}{c_2} - t \right) \right]$$

where

$$c_2 = \frac{\alpha_w}{\sin \theta} .$$

The boundary condition at the interface is

$$u_z \bigg|_{z=0} = s_z \bigg|_{x=0}$$

which gives the plane wave refraction formula:

$$\frac{\sin \phi}{\sin \theta} = \frac{\alpha_w}{\alpha_r \text{ or } \beta_r} . \quad (\theta = \text{angle of incidence of water wave})$$

and

$$u_z = a \cos \theta$$

$$a = \frac{u_z}{\cos \theta} .$$

Hence a vertical component of ground motion of amplitude $u_z$ gives rise to a pressure wave of amplitude

$$p_0 = \frac{\rho \omega}{\cos \theta} u_z .$$

Taking

$$\alpha_r = \sqrt{3} \beta_r = 6.0 \text{ km s}^{-1}$$

$$\alpha_w = 1.5 \text{ km s}^{-1} ,$$

$\theta \text{ max } \approx 15^\circ$ for incident $P$-waves

$\approx 23^\circ$ for incident $S$-waves

(for $\phi = 90^\circ$)
At a frequency of 20 Hz, we find
\[ p \approx 2U_z \]
where \( p \) is the amplitude in microbars of the pressure wave produced by a ground motion of amplitude \( U_z \) m\( \mu \).

For very close earthquake magnitude determination we use a formula similar to that given by Brune & Allen (1967)
\[ M = \log U_z + \log G_{wa} - \log A_0 + A_{20} \]
where
\( U_z = \) ground motion amplitude in millimetres,
\( G_{wa} = \) gain of Wood-Anderson torsion seismometer,
\( A_0 = \) distance correction factor as given by Richter (1958),
\( A_{20} = \) Additional amplitude/distance correction factor determined by Brune & Allen (1967) for 20 Hz seismic waves.

We chose 20 Hz as the frequency because both theory and experiment indicate that for nearby microearthquakes the highest signal-to-noise ratio is obtained in this region and on the visible records this is a predominant frequency.

The magnitude formula then becomes,
\[ M = \log p_{20} (\mu \text{ bar}) - 2.85 - \log A_0 + \log A_{20}. \]

For a magnitude 0 earthquake at a distance of 10 km, this would give a pressure amplitude at 20 Hz of about 2.5 m\( \mu \). The ambient ocean noise in the bandwidth 5–100 Hz is of the order of 1 m\( \mu \) (Wenz 1962), so that a magnitude 0 event at 10 km distance is probably at the lower limit of detectability. At closer ranges, events of magnitude below 0 may be detected, and we believe we have recorded numerous swarm events of magnitude less than 0.

At distances beyond about 10 km, the 20 Hz seismic waves are attenuated rapidly, and lower frequencies predominate. For these it is necessary to estimate an attenuation factor for the observed predominant frequency, and to allow for the frequency dependence of the pressure/amplitude relation.

During ‘Hypogene’ expedition two small (\( M \approx 1.8 \) and \( M \approx 1.5 \)) events have been identified both on hydrophone records and on a nearby (\( \Delta \approx 60 \text{ km} \)) land station (Guaymas). The magnitudes estimated from both observations are in reasonably good agreement.

As a comparison with microearthquake studies on land, we note that an ambient sea-noise level of 1 m\( \mu \) is equivalent to 20 Hz to a seismic noise level of 0.5 m\( \mu \). Brune & Allen (1957) found that quiet sites in Southern California typically had a noise level of 0.05 m\( \mu \). In general, then, microearthquake studies at sea are limited by a noise level about an order of magnitude greater than that for quiet sites on land; this is still sufficiently low that useful microearthquake studies may be done with hydrophones.

3. Frequency dependence

To determine the optimum overall frequency response of a recording system using hydrophones, we compare in Fig. 1 a typical ambient sea noise spectrum (Wenz 1962) with that of a small (\( M \approx 0 \)) earthquake at a distance of about 10 km. The signal:noise ratio is greatest at around 20 Hz. The system on ‘Hypogene’ used no filtering
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FIG. 1. Typical Earthquake Pressure Spectrum. Long dashes, spectrum in the absence of sea surface. Short dashes, spectrum when interference due to reflections from sea surface are taken into account. Solid line, typical ambient noise spectrum from Wenz (1962). Spectra are relative to base level of $3 \times 10^{-5}$ microbars.

FIG. 2. Frequency response curve of sonobuoys in db re. 1 μbar at 440 Hz. Strip chart recorder cut-off at 100 Hz, limits high frequencies in recorded signal.
except that in the sonobuoy amplifier (a high-pass system), and the upper cut-off of about 100 Hz set by the mechanics of the recording system. The approximate response curve is shown in Fig. 2.

Another factor to be considered in the frequency response is the effect of the wave reflected from the ocean surface. The hydrophone depth should be selected so as to avoid destructive interference between the incident and surface reflected waves at the predominant frequencies. The pressure at a depth $d$ due to the sum of an incident wave of length $\lambda$ and its surface reflection is, apart from a phase factor

$$p = p_I + p_R = 2p_1 \sin \frac{2\pi d}{\lambda}.$$

For a frequency of 20 Hz, the first pressure antinode is at a depth of approximately 20 m. The hydrophone depths most commonly used are 20 and 100 m. The above correction factor should be applied to the observed pressure $p$ in order to calculate the amplitude $p_1$ of the incident wave. The assumption $p_I = p$ will cause little error in the estimation of earthquake magnitudes.

4. Instrumentation and recording equipment

**Sonobuoy/Hydrophone system**

Telemetering radio-sonobuoys have been successfully used for many years in seismic refraction studies at sea, and the proven ability of modified sonobuoys to detect weak mantle refracted arrivals was one of the factors which encouraged us to attempt earthquake detection and location with sonobuoys. The sonobuoy/hydrophone system converts seismically-generated pressure fluctuations in the ocean into a radio signal which is transmitted to the receiving apparatus. The hydrophone system is suspended by a loose-coupled (to reduce noise) cable from the floating sonobuoy at a depth of between 20 and 100 m. Pressure fluctuations are converted by one or more piezo-electric crystals into weak electrical signals which are pre-amplified in the hydrophone system and transmitted up the hydrophone cable to the sonobuoy, where the signal is further amplified, filtered, and used to frequency-modulate a VHF radio signal transmitted by the sonobuoy.

We desire a system which is reasonably well calibrated and linear, and which has an overall gain sufficiently great that the signal detection threshold is limited by the ambient sea noise and not by the sensitivity of the system. The frequency response should be such as to give the best overall signal-to-noise ratio for events of interest. A discussion on optimum frequency response was given in the preceding section. Adjustment of overall system gain and filtering can of course be done between receiver and recorder, but it is best if as much amplification and filtering as possible is done at the signal source.

For most purposes standard naval expendable sonobuoys are adequate, and we have used these, largely because of their operational simplicity and the great savings in ship time gained by their use. The frequencies of greatest interest in local earthquake studies are about 20 Hz, and normally it is not necessary to modify the frequency response of the sonobuoys. The sensitivity of these instruments is such that the signal produced by the ambient sea noise is well above the electrical and mechanical noise level of the receiving and recording system, and is therefore adequate. One disadvantage of expendable sonobuoys is that the calibration is in general not very accurate. In addition, most sonobuoys contain an automatic gain control to match the gain to the ambient noise level. This can be disabled, if desired. However, comparative magnitudes and sea noise levels are reasonably consistent, and we believe we know the calibration to within a factor of two. We have assumed the automatic gain control had no effect. We will improve the system calibration in future work.
Receiving and recording system

The receiving and recording station may be on board a ship or aircraft or, in some cases, on land. The radio signals from the sonobuoys are received and recorded, either visually or on magnetic tape. On board R/V *Melville* during the 'Hypogene' Expedition in the Gulf of California, the radio signals were picked up by three highly directional VHF antennas on the ship's masts, amplified, and fed to a set of VHF FM radio receivers in the ship's operations room. The receiving antennas were rotormounted, and their orientations remote-controlled from the operations room, where the signals were monitored. Each radio receiver was tuned to the radio frequency of one of the sonobuoys deployed at the time, and the output from each radio was fed into a channel of a multichannel strip-chart recorder. Recording speeds varied between 1/2 mm s\(^{-1}\) during most of the month-long recording period, to 50 mm s\(^{-1}\) during periods of intense earthquake swarm activity. For shipboard operation the multichannel strip-chart recorder has several advantages: compactness, flexibility in gain and recording speed, and ease of monitoring several channels simultaneously.

The radio receivers were modified to give DC response. This was extremely useful for tuning into weak signals, since with an FM system a slight detuning of the radio produces a DC output which verifies that the receiver is locked into the signal. This also served as a valuable method of system calibration; by observing the recorder trace deflection produced by detuning the radio a given amount from the carrier frequency, a direct relation between trace deflection and frequency shift is found. This can be used in conjunction with the characteristic pressure/frequency shift relation of the sonobuoy to obtain the overall gain.

5. Operational procedure

During most of the 'Hypogene' Expedition, the emphasis was on detailed bathymetric, heat flow and other studies, which took priority over the seismicity studies. Fortunately the areas being studied, mainly the Gulf basins, are also areas of high seismicity. Hence we could simply launch expendable sonobuoys at intervals during the surveying operation, and record these for as long as they were within range and continued transmitting. The most efficient use of the available sonobuoys to obtain reasonably complete coverage of a given area requires some prior knowledge of the features which are most likely to be seismically active, since the high frequency seismic waves to which the hydrophone is most sensitive attenuate rapidly and small events can be detected only if they are within a few tens of kilometres of the sonobuoy. On 'Hypogene' we used two types of sonobuoy, one with a 3-day lifetime, and one with a 3-hr lifetime. Usually the 3-day instruments were used when in a basin or other region where the ship was expected to remain within range for an extended period, and the 3-hr instruments when the ship was steaming a more or less straight course, and hence would be out of range within a few hours. We found that we could continuously monitor seismicity while not interfering with the other surveying operations.

During a later part of the cruise, the seismic work had priority. This was clearly a great advantage, since it enabled us to place the sonobuoy arrays in desired positions and record for extended periods. This advantage was reflected in the improved results obtained during this period; more earthquakes were recorded and far greater accuracy of location was possible.

Whenever necessary we used the seismic profiler airgun to obtain the position of the drifting sonobuoys. The time delay between the airgun firing pulse and the arrival of the direct water wave or that reflected off the bottom at the sonobuoy gives the ship-sonobuoy distance. Repetition of this procedure from several different ship positions allows location of the sonobuoy by triangulation. Where the drift rate is large, and the sonobuoys have been drifting a long time, this procedure is vital for accurate earthquake locations. Airgun ranging was particularly useful during a swarm
of small earthquakes occurring in Guaymas Basin on the morning of April 11. When it was seen that the swarm was occurring (on a sonobuoy launched a few hours earlier) additional sonobuoys were launched in a close (\(\sim 5\) km) array around the estimated epicentre. The system of simultaneous recording on the strip-chart recorder was particularly valuable at this time, as it allowed immediate first estimates of epicentre locations, and this helped in the deployment of the rest of the array. Subsequently the ship steamed a pattern around the array, and carried out airgun ranging at frequent intervals. Precise navigation is essential; navigation on 'Hypogene' was by satellite, supplemented by frequent radar fixes. With a sonobuoy array consisting of several sonobuoys directly over the epicentre the ultimate location accuracy will be limited by the accuracy with which the floating sonobuoys can be tracked and this is ultimately limited by navigation accuracy.

In the future we plan to use aircraft for launching and recording the sonobuoys. The advantage in speed thus obtained may be crucial in aftershock and swarm studies. The few hours of recording obtainable with an aircraft may yield valuable data if the level of activity is high. We have successfully carried out tests with a DC-3 aircraft, and hope to use this for studies in case of a major earthquake or swarm within flying distance of San Diego.

6. Moored arrays

The discussion so far has been concerned with free-drifting expendable sonobuoys. An alternative approach is to use anchored sonobuoys, designed for a longer lifetime (several days). This procedure has the advantage that the sonobuoy position is fixed. Furthermore, if the region studied is within radio distance of the shore (100–150 km if the shore station can be set up on high ground) it is possible to deploy the sonobuoys, return to record on shore, and return to retrieve the sonobuoys only at the end of the recording operation, thus saving ship time. Moored sonobuoys are intrinsically noisier. They also have to be specially built, and the necessary anchors, cable, floats and marking devices provided. This increases the cost of operation. Deep-sea moorings can be made reliable only at great cost. Mooring is important in cases where no continuous recording ship is available.

During 'Hypogene' moored arrays were set up near Guaymas and Topolobampo with receiving stations nearby on land. The recording equipment was the same as that used aboard ship, except that recording was done using portable smoked-paper seismographs. The Topolobampo station had to be abandoned due to radio interference, but some valuable earthquake recordings were made at the Guaymas station. Details of these results are discussed in a separate paper (Brune et al. 1973).

7. Preliminary results

Interpretation of hydrophone records

Figs 3–5 give examples of hydrophone records of small earthquakes. Examples 3 and 4 were taken using a ship-operated Sanborn strip-chart recorder and example 5 was obtained at Guaymas using land-operated smoked paper recorders of the type described by Prothero & Brune (1971).

On these records the onsets of the events are sharp and the predominant frequencies are about 20 Hz. Several phases are evident and these have been identified in the figures. \(P\) and \(S\) refer to arrivals corresponding to \(P\)- and \(S\)-waves travelling through the crust and being converted to pressure waves under the hydrophone. The subscripts on the \(P\) symbols indicate the number of bottom reflections the wave has undergone before arriving at the hydrophone. \(P\) refers to the first arriving pressure wave; \(P_1\) refers to a similar wave which has undergone one additional bottom reflection.
Fig. 3. Event recorded aboard ship with free floating sonobuoys.
FIG. 4. Several events of the Guaymas swarm of 1972 April 11 recorded aboard ship.
FIG. 5(a) and (b). Earthquakes recorded at the land receiving station, Guaymas, Mexico. Events were detected on free floating hydrophones.
The symbol $T$ refers to the $T$ phase, i.e. the wave travelling most of its path as a pressure wave in the water layer, with a velocity of about $1.5 \text{ km s}^{-1}$. $T$ phases from earthquakes have been observed on ocean bottom seismographs (Nagumo et al. 1970). The $T$ phase is not observed on very close events and only becomes clear when the distance is great enough for the slow travelling pressure wave in the water to separate from the faster travelling $P$- and $S$-waves in the crust.

Interpretations are not always obvious and the interpretations we have made are tentative. The $T$ phase is identified primarily by its low velocity and the reflected phases are identified primarily by the interval between successive reflections, and correlation with water depth. The amplitudes of the reflected phases decrease rapidly with distance whereas the amplitude of the $T$ phase (relative to the $P$ and $S$ phases) increases with distance. This is due to the low intrinsic attenuation of pressure waves

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**Fig. 6.** Schematic diagram of Gulf of California showing locations of earthquakes and earthquake swarms.
in water (high $Q$) and the fact that the pressure waves are trapped in the water channel whereas the $P$- and $S$-waves may be radiating into deeper layers.

**Locations**

The problem of locating earthquakes with an array of near-surface hydrophones is similar to that of locating local earthquakes on land, with a few differences. The relative delay time in the water must be taken into account. Because of the velocity contrast between the water and the crustal rocks, the ray paths of $P$-waves in the water are nearly vertical, and hence the delay time can be estimated from the depth of water under the receiving hydrophone. The correction for water depth can be a serious problem if the bathymetry is not accurately known. A second difference between hydrophone locations and land locations is the weakness of the $S$-wave, especially at short distances. The $S$-wave is often mixed with multiple reflections. The lack of a clear $S$-wave can be a serious problem in cases where locations are attempted with only a few stations.

The locations of many events recorded here are considerably more precise than is possible for teleseismic location in the Gulf (20–50 km uncertainty, see Thatcher & Brune 1971). Fig. 6 and Table 1 summarize the results of our observation of seismicity in the Gulf.

The Gulf of California basins are believed to be areas of crustal spreading linked by a series of en-echelon transform faults. It appears that all the swarms we observed were associated with spreading centres, rather than transform faults. This result is consistent with Sykes (1970) results for swarms on the Mid-Atlantic Ridge. The spreading centres had a remarkably high seismicity. Individual earthquakes were

![Fig. 7. Ship track at Guaymas Basin. All times are GMT. Open circles are positions of sonobuoy drops. X's are positions of airgun shots used in sonobuoy relocation.](https://academic.oup.com/gji/article-abstract/34/3/365/631320/fig07)
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Table 1

Episodes of activity observed during Hypogene (see also Fig. 1)

<table>
<thead>
<tr>
<th>Date</th>
<th>Event Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1972 March 16.</td>
<td>Swarm-type sequence believed from South Pescadero Basin (although an ambiguity in location exists). Nine events of magnitude 1-3 in a 7-hr period.</td>
</tr>
<tr>
<td>1972 March 21-24.</td>
<td>Swarm-type earthquake sequence from the south-west corner of Guaymas Basin. Ten to twelve events, magnitude in range 0-2 in a 50-hr period.</td>
</tr>
<tr>
<td>1972 April 2-3.</td>
<td>Sequence of earthquakes near the mid-point of Farallon Basin. Ten to twenty small events during the 24-hr recording period, with a small swarm towards the end.</td>
</tr>
<tr>
<td>1972 April 5-6.</td>
<td>Continuation of earthquakes from the central region of Farallon Basin. Twelve events during 20-hr recording period (previous recording was interrupted by the necessity of steaming to La Paz for ship repairs).</td>
</tr>
<tr>
<td>1972 April 6-8.</td>
<td>Approximately 15 events, near deepest part of North Pescadero Basin during 40-hr recording period.</td>
</tr>
<tr>
<td>1972 April 9.</td>
<td>Earthquake swarm; 50 small events in the central part of Farallon Basin over a 10-hr period, with most of the activity occurring over a short time near the middle of this period.</td>
</tr>
<tr>
<td>1972 April 11.</td>
<td>Swarm of about a thousand recorded events from Guaymas Basin, lasting 6 hr. We expect to be able to locate this swarm with great accuracy on further analysis (including accurate depths) and correlate this with the basin topography. This is almost certainly the most detailed study ever obtained of this type of event, and is expected to provide much valuable information on the nature of earthquake swarms.</td>
</tr>
</tbody>
</table>

also recorded at various places as indicated in Fig. 6. Some recording was done farther south, over the East Pacific Rise, but no earthquakes were recorded. The recording time was not sufficient to conclude, however, that the seismicity was less than in the Gulf.

During the Guaymas swarm we deployed enough instruments to get relatively accurate locations. Fig. 7 shows the ship track during deployment of the array and during subsequent recording hours. The ship track was determined using several satellite fixes and numerous radar fixes from targets on both sides of the Gulf. The bathymetric and airgun profiling was carried out continuously and, along with data from other tracks during Hypogene, provided a considerably detailed picture of the bathymetry and structure of the Guaymas spreading centres.

The data necessary for the location procedure are given in Table 2. This data consists of a series of locations outlining the ship course plus ranges to the various hydrophones determined from the water wave travel time. Fig. 8 shows the drift

Table 2

Data used in relocation of sonobuoys

<table>
<thead>
<tr>
<th>Shot time</th>
<th>Ship position</th>
<th>Sonobuoy number</th>
<th>Direct arrival</th>
<th>Sediment reflection arrival (s)</th>
<th>Distance (NM)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1704</td>
<td>27° 20' 0 N</td>
<td>85</td>
<td>nv</td>
<td>7.95</td>
<td>6.09</td>
</tr>
<tr>
<td></td>
<td>111° 25.2 W</td>
<td>89</td>
<td>nv</td>
<td>9.0</td>
<td>7.01</td>
</tr>
<tr>
<td>1807</td>
<td>27° 17.8 N</td>
<td>85</td>
<td>nv</td>
<td>3.6</td>
<td>2.19</td>
</tr>
<tr>
<td></td>
<td>111° 29.9 W</td>
<td>89</td>
<td>nv</td>
<td>3.8</td>
<td>2.36</td>
</tr>
<tr>
<td>1829</td>
<td>27° 17.0 N</td>
<td>85</td>
<td>nv</td>
<td>3.7</td>
<td>2.17</td>
</tr>
<tr>
<td></td>
<td>111° 32.1 W</td>
<td>89</td>
<td>1.25</td>
<td>2.8</td>
<td>1.05</td>
</tr>
<tr>
<td>1838</td>
<td>27° 16.7 N</td>
<td>85</td>
<td>nv</td>
<td>4</td>
<td>2.6</td>
</tr>
<tr>
<td></td>
<td>111° 33.0 W</td>
<td>89</td>
<td>1.5</td>
<td>2.8</td>
<td>1.4</td>
</tr>
</tbody>
</table>
tracks of each sonobuoy as derived from the launch position and the airgun triangulation.

Table 3 lists the arrival time and sonobuoy position data for one well-recorded event which was located. The small arcs struck in Fig. 9 indicate the consistency of the data used to make the locations. As can be seen from Fig. 1 the location accuracy is about 2 km and is sufficient to establish that the epicentres are in the 2 km wide graben associated with the Guaymas spreading centre.

Table 3

<table>
<thead>
<tr>
<th>Sonobuoy No.</th>
<th>Arrival time</th>
<th>Sonobuoy location</th>
</tr>
</thead>
<tbody>
<tr>
<td>85</td>
<td>1451:5:0</td>
<td>27° 16:9 N</td>
</tr>
<tr>
<td></td>
<td></td>
<td>111° 30:37 W</td>
</tr>
<tr>
<td>87</td>
<td>1451:5:8</td>
<td>27° 15:9 N</td>
</tr>
<tr>
<td></td>
<td></td>
<td>111° 27:5 W</td>
</tr>
<tr>
<td>88</td>
<td>1451:5:2</td>
<td>27° 15:2 N</td>
</tr>
<tr>
<td></td>
<td></td>
<td>111° 31:7 W</td>
</tr>
<tr>
<td>89</td>
<td>1451:5:0</td>
<td>27° 17:2 N</td>
</tr>
<tr>
<td></td>
<td></td>
<td>111° 32:0 W</td>
</tr>
</tbody>
</table>
Seismicity

We have assigned magnitudes using the magnitude formula given above. The largest events recorded occurred near South Pescadero Basin on March 16 and had a magnitude of about 3. Several events with magnitudes 2 to 3 were recorded while several hundred events were recorded with magnitudes between < 0 and 2. The 'b' value, the negative slope of logarithm of the cumulative number vs. magnitude curve (Richter 1958, p. 359) was approximately 1, in agreement with observations along the San Andreas fault on land (Allen et al. 1965). This was approximately true for the sample of earthquakes with the three major swarms omitted, as well as for the swarms themselves and for the total sample.

Nevertheless, there is an obvious difference in the pattern of occurrence of earthquakes we have observed from that observed for microearthquakes on land (Brune & Allen 1967). A far greater proportion of events recorded here occurred in swarms rather than as random individual events. The typical pattern of seismicity observed near Anza, California by Brune and Allen was a more or less continuous microearthquake activity of about 15 events per day for various periods covering more than a year, whereas in the Gulf the pattern was one of general quiescence interrupted by swarms of events with durations from a few hours to a few days. If the three major swarms were excluded, the overall rate of seismicity observed in this study was about three events per day of noise-free recording time. Including the swarms gives an
overall rate of about 40 events per day, primarily as a result of the nearly 1000 events recorded during 6 hr in the Guaymas Basin. In comparison the seismicity observed in the microearthquake survey along the San Andreas Fault (Brune & Allen 1967) varied from less than one event per day over much of the locked portion of the fault to about 10–20 events per day along the active parts of the San Jacinto fault. Since the equivalent background noise level in our studies is about an order of magnitude higher than for the study of Brune & Allen we conclude that the overall seismic activity in the Gulf, exclusive of the three major swarms, is comparable to that of active parts of the San Jacinto fault. If the events in the swarms are included the rate of seismicity in the Gulf is far greater than even the most active parts of the San Jacinto fault.

The events recorded here are all much smaller than the events of the northern Gulf of California swarm of 1969 (Thatcher & Brune 1971). The events studied by Thatcher and Brune ranged in magnitude up to 5.7 and most of the events were of magnitude greater than 4. The largest event in the Guaymas swarm was about magnitude 2.

Conclusions

Telemetering sonobuoys have been successfully used to study microearthquake seismicity in the Gulf of California during the ‘Hypogene’ Expedition of the R/V Melville in March and April of 1972. P, S, T and multiply reflected P phases have been identified. The onsets of the P phases are sharp and thus may be used for accurate locations. In the 6 weeks of operation (two of which were given high priority for hydrophone operation) we recorded numerous individual earthquakes, several earthquake swarms of 5–20 events and one swarm in the Guaymas Basin consisting of several hundred events in a period of 6 hr. During the Guaymas swarm an array of hydrophones was deployed over the hypocentral region allowing location accuracies of about 2 km. The epicentre of the swarm was in a graben associated with the spreading of the Guaymas Basin. During the 6 weeks of operation almost every basin produced a swarm or swarm-like sequence of events. The overall microearthquake activity, including swarms, is higher than observed near the San Andreas fault on land.

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