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# A POSSIBLE SHOCK EFFECT ASSOCIATED WITH SEAQUAKES

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## A POSSIBLE SHOCK EFFECT ASSOCIATED WITH SEAQUAKES

### INTRODUCTION

The term 'sequake' has been defined to mean simply an earthquake whose epicenter lies at sea rather than on land. Initially, the word was used to describe any seismic disturbance of the ocean including tsunamis, but the term has now been restricted to refer to only the effects of earthquakes felt on board vessels at sea (Richter, 1958). While there is little difference between the cause of an earthquake at sea and that of any comparable continental earthquake, there exist considerable differences in effect.

Descriptions of seaquakes have remarkable similarities. The first such were presented by Rudolph in his papers "Ueber Submarine Erdbeben und Eruptionen" published in 1887 and 1895, in which he gives over 400 descriptions from ships' logs. At the time he performed his studies, comparatively few of the shocks felt on board ship had been recorded by seismic instruments on land. It was originally thought that these shocks were probably due to small earthquakes. Improved instrumentation has since shown that most of the earthquakes were probably of moderate size (Gutenberg and Richter, 1954).

In the ship reports of seaquakes collected by Rudolph, although many are of a minor nature, there are those that are not minor. A seaquake experienced by several ships near the Cape Verde Islands in 1833 was described as "a succession of heavy shocks felt, cracking sounds, ship trembled most violently. Vibration terrible." Another was described by Captain John Tadot of 'H. M. S. Sylvia' in the North Pacific near Japan in 1870 as "Three distinct shocks of an earthquake. Three bumps as if the ship had grounded and had been lifted up and left again by the sea." Captain Garden of the 'Northern Monarch' in 1878 at a position in the Indian Ocean 12°4' S, 84° 38' E wrote: "Observed sea throw up to a great height possibly 80 feet or more in a column; this occurred three or four times, each upheaval lower than the preceding one; effect similar to that produced by a torpedo. Examining the position immediately from aloft, the water at the place was observed to break three or four times like heavy breakers and then smooth down and we saw no more. The place from the ship about S. E. by E. distant 5 to 6 miles."

Some earthquakes have produced measured results that are almost unbelievable. In the instance of the Kwanto Earthquake which occurred on September 1, 1923 in Japan, a hydrographic survey of Sagami Bay shortly afterward revealed permanent peak-to-peak vertical changes in the floor of the bay amounting to 1500 feet, (Heck, 1965).

Descriptions of ordinary seaquakes are distinguished by the following features:

- a. Most report a loud rasping or grating sound sometimes accompanied by violent shaking as if the ship had hit a hard object and was skidding over it.
- b. Some report at least one (Eiby, 1957) and sometimes two or more sharp jolts against the hull of the ship.
- c. Many are accompanied by sounds variously described from 'distant thunder' to 'a tremendous and explosive noise.'
- d. The energy in some shocks is sufficient to cause large ships at distances of over 100km from the epicenter to list from 5 to 10 degrees.
- e. Many ships are in deep water, i. e.,  $> 1000$  m depth when the seaquake is experienced.
- f. Usually there is no apparent motion of the surface of the water connected with the shock.

Let us consider how the observed effects of seaquakes may be brought about. If we regard much of the earth as an elastic solid, the sudden release of energy during an earthquake generates two types of elastic waves: compression waves or P waves and shear or S waves; the designations P and S refer to the primary and secondary arrivals at the surface. The compression wave, being faster, arrives first and is transmitted from the mantle, through the crust to the floor of the ocean. Since water is a much less dense medium than rock, the wave is refracted upward and would account for the description of a jolt to a ship. The S wave does not propagate through the water, but rather is reflected at the rock/ocean interface and is partially converted to a compression wave. This would then propagate upward and would account for a second jolt to a ship of less intensity than the first and distinctly separated in time from it. Additional shocks are caused by secondary P waves, reflected waves, etc. Sounds accompanying the seaquakes have a variety of causes. Some are the direct result of the shock waves striking the ship. Others can be attributed to the transfer of the shock into the air; e.g. the sound of distant thunder can be explained by the refraction of sound waves in atmospheric temperature gradients.

With the exception of the studies performed by Rudolph, effects of seaquakes have here-to-fore received little attention. This may be due to the lack of significant damage done to shipping or to the fact that 'in situ' measurements are not commonplace. In either case, of the ships that have experienced a large

seaquake, most have been well outside (> 100km) the epicentral area. The relationship between earthquake belts and shipping density is given in Figure 1.

#### DATA

On April 29, 1970 at 1401 GMT a large earthquake of shallow focus and registering 7.5 on the Richter scale occurred in the Guatemala Basin off the coast of the Mexican state of Chiapas (CFSLP, NOAA). The location of the epicenter was 13.5°N, 92.5°W. The main shock was preceded by a strong foreshock at 1122 GMT, which reached a magnitude of 6.5R. Between 1122 GMT and 2015 GMT over 70 seismic events were recorded; there were no reports of damage nor of fissuring. The mareographic station at Salina Cruz, Oaxaca, was out of order and no report is available on the tide.

At 1140 GMT, approximately 18 minutes after the initial foreshock, the 11.5μ IR radiometer on board theITOS-1 spacecraft recorded the thermal information presented in Figure 2. A circular area approximately 60km in diameter in the immediate vicinity of the earthquake epicenter shows an anomalous temperature enhancement of approximately +3°K over the surrounding environment. An average water temperature of 302°K is indicated by the IR data for this area. It is of some interest to note that this is slightly lower than that (86°F or 303°K) given by charts of mean sea surface temperature (Figure 3) for April 1970 prepared by the U.S. Navy Oceanographic Office. Other measurements of sea surface temperature by the radiometer at this time, however, are somewhat lower still than the corresponding 'in situ' measurements. This is due to the radiative transfer of the atmosphere, the effect of which can be determined from the radiance.

The radiance (I) as measured by the s/c radiometer can be approximated by the equation

$$I = \tau_s B(T_s) + (1 - \tau_s) B \bar{T} \quad (\text{Kunde})$$

which assumes a unit level atmosphere where

B = Planck intensity [@ 11.5μ]

T<sub>s</sub> = Surface Temperature

τ<sub>s</sub> = Transmissivity of atmosphere

$\bar{T}$  = Atmospheric Temperature

For the case  $T_s = 303^\circ\text{K}$ ,  $\bar{T} = 290^\circ\text{K}$ ,  $\tau_s = 0.6$  the temperature as measured by the spacecraft sensor would be  $298^\circ\text{K}$  or about  $5^\circ\text{K}$  lower than the actual surface temperature. The presence of any clouds in the field of view of the sensor would serve to lower the radiance even more.

The  $3^\circ\text{K}$  temperature anomaly is not observable in ITOS-1 or Nimbus 4 data in other orbits prior to or following the earthquake.

On August 11, 1970 at 1020 GMT a major shallow focus earthquake of Richter magnitude 7.6 occurred in the region of the New Hebrides Islands in the South Pacific (CFSLP, NOAA). At 1255 GMT the area was observed by the  $11.5\mu\text{IR}$  radiometer on board the Nimbus 4 spacecraft. The water temperature in the immediate vicinity of the earthquake, i.e.,  $14.1^\circ\text{S}$ ,  $166.7^\circ\text{E}$ , was recorded to be between  $296^\circ$  and  $297^\circ\text{K}$  or about  $2^\circ$  warmer than the surrounding area.

In both the case of the Guatemala Basin seaquake and that of the New Hebrides quake it is postulated that the observed temperature effects could be the direct result of a shock wave caused by the quake and propagating through water and atmosphere.

## DISCUSSION

The {shock, sound, elastic, compression} wave resulting from the release of {stress, strain} energy by an earthquake whose hypocenter lies in the upper mantle travels through the upper mantle at speeds of from 7 to 8 km/sec (Richter, 1958). Between the mantle and the crust is a sharp, almost worldwide discontinuity known as the Mohorovicic discontinuity or 'Moho'. Above this is the crust, a layered structure of decreasing seismic velocities, the uppermost portion of which is composed of sediments with very low velocities and very high attenuation (Stacey, 1969). In instances where it has been measured, there is a direct correlation between depth and displacement. In the case of the Idu (Japan) earthquake, the measured displacement at the surface was three feet, whereas in the tunnel of Tama on the Tokyo-Kobe line 530 feet below, the displacement was 8 feet (Tazieff, 1964).

The average thickness of the continental crust is between 30 and 40 km while the average thickness of the oceanic crust is only 5 km and of much simpler and more uniform basaltic structure (Stacey, 1969). A shock wave being propagated upward from a continental hypocenter at a depth of 30 km would be severely attenuated and fractionated by the crustal structure and in particular by the inhomogeneous fractures near the surface. In the case of a seaquake, propagation is through a thin crust followed by the homogeneous medium of water. Thus, less attenuation would be expected.

The compression wave from the earthquake propagates as a shock wave through the water. Since the piston motion of the ocean floor is finite, the shock wave must be followed by a rarefaction.

The rarefaction wave velocity is the sound velocity behind the shock plus the particle velocity ( $u$ ). A rarefaction wave will therefore overtake the shock front and, by cooling and expanding the fluid behind the front, will cause a rapid weakening of the shock. It can be readily shown, then, that shock propagation is a critical function of the relationship between the piston like motion of the ocean floor and the depth. If the initial compression wave has a duration of 0.1 sec the rarefaction wave would overtake it at a depth of less than 500 meters. (This is illustrated by the characteristic diagram for a 6 kilobar shock given in Figure 5.) If the total water depth is  $\sim 2500$  meters, as in the case of the Guatemala Basin earthquake, any effect upon the surface would be small for reasonable displacements of the ocean floor. A value  $t > 0.25$  sec is required to produce the effect seen in the IR data. This would require a peak deflection of the ocean bottom over the epicenter in excess of 70 meters. The water at the surface would not necessarily break since the height/width (60km) ratio would only be on the order of 0.001; however, the effect upon an object in the immediate vicinity of the center of such a shock would be catastrophic.\* As one moves away from the epicenter, the shock strength would rapidly diminish due to both the dissipation of the shock front and the effects of rarefaction waves.

When a shock wave moves with a velocity  $U$  through a stationary fluid of initial pressure  $P_0$  and specific volume  $v_0$ , the pressure  $P$ , specific volume  $v$ , and particle velocity  $u$ , of the fluid behind the shock are determined by the Rankine-Hugoniot conditions (Rankine, 1870, Hugoniot, 1887). These express the conservation of mass, energy, and momentum of an element of the fluid through which the shock front moves and are given by:

$$u = [(p - p_0)(v_0 - v)]^{1/2}$$

$$U = v_0 [(p - p_0)/(v_0 - v)]^{1/2}$$

$$\Delta H = (1/2)(p - p_0)(v + v_0)$$

where  $\Delta H$  is the specific enthalpy (defined as  $\Sigma$  internal energy per gram +  $pv$ ) increment of an element of fluid when it passes through the shock front (Richardson, 1947). In addition to the above relationships one also needs the

\*The characteristic time  $O_p$ , is the time for a plate to deflect to a fraction  $1 - 1/e = 0.63$  of its final set in response to a step pressure and is called the plastic time by Kirkwood.  $O_p$  for a diaphragm 10.0 inches in diameter and 0.1 inch thick with a yield stress of 60,000 p.s.i. is 600  $\mu$ sec, (Cole, 1948).

speed of sound in the fluid  $c$ , the Riemann function  $\sigma$ , the undissipated enthalpy  $\omega$ , and the equation of state of water.

$$c = (\partial p / \partial \rho)^{1/2}_s \quad \rho = 1/v, \quad s = \text{entropy}$$

$$\sigma = \int_{p_0}^p [v(p', S)/c(p', S)] dp'$$

$$\omega = \int_{p_0}^p v(p', S) dp'$$

Tait equation of state:

$$v(0, T) - v(p, T)/v(0, T) = (1/n) \log [1 + p/B(t)]$$

where  $t = (T - 273.16)^\circ\text{C}$ ;  $n$  and  $B(t)$  are empirically determined.

The solutions of the above equations which give the desired variables  $u$ ,  $U$ ,  $c$ ,  $v$ , and  $T$  as functions of  $p$  involve a series of successive approximations and have been carefully dealt with by many authors. A comparison of results is given in Table 1 for two pressures, 5 kb and 10 kb.

Table 1  
Initial Temperature  $T_0 = 20^\circ\text{C}$ ,  $P_0 = 1$  atmosphere

	$p$ (kb)	$u$ (m/sec)	$U$ (m/sec)	$c$ (m/sec)	$v$ (cc/gm)	$T$ ( $^\circ\text{C}$ )
Richardson	5	251.0	1975.0	2230.0	0.8668	—
Sternberg	5	248.8	2013.7	2264.5	0.8781	32.26*
Rice	5	250.0	2008.0	2350.0	0.8773	36.0
Kirkwood	5	—	—	—	—	35.7
Arons**	5	240.2	2038.9	—	—	—
Richardson	10	425.5	2335.0	2755.0	0.8120	—
Sternberg	10	425.1	2356.6	2707.9	0.8211	50.48*
Rice	10	426.0	2352.0	2800.0	0.8204	54.0
Kirkwood	10	—	—	—	—	55.5
Arons**	10	408.9	2395.6	—	—	—

\*calculated for salt water

\*\* $T_0 = 25^\circ\text{C}$



Using data supplied by the United States Naval Ordnance Laboratory (Sternberg, 1972) for explosions in salt water, the pressure/temperature states of a fluid immediately behind a shock front can be plotted on a curve, commonly referred to as a "Hugoniot" (Figure 4). An element of fluid initially at a state  $(p_0, T_0)$  reaches a state of  $(p_1, T_1)$  immediately behind the shock and then returns to a state  $(p_0, T_0')$  along the adiabatic (isentropic) curve.

In Figure 4,  $P_0 = 1 \text{ bar}$ ,  $T_0 = 20^\circ\text{C}$ . To raise the temperature of the water at the surface from  $T_0 = 20^\circ\text{C}$  so that  $22^\circ\text{C} < T_0' < 23^\circ\text{C}$ , would require a surface pressure  $p_1$  at the ocean/air interface of between 6 and 8 kilobars, or  $\sim 100,000 \text{ p.s.i.}$  ( $14513 \text{ p.s.i.} = 1 \text{ kilobar}$ ). This pressure is probably higher than is necessary to account for the thermal anomaly observed by the spacecraft sensors since it assumes the full temperature rise is in the water and neglects the effect of the propagation of the shock into the atmosphere.

For a 6 kilobar shock  $c = 2358.43 \text{ m sec}^{-1}$  and substitution of the corresponding values of

$$p_0 = 1 \text{ bar} = 1.01355 \text{ nm}^{-2}$$

$$v_0 = 1.0134 \times 10^{-3} \text{ m}^3 \text{ kg}^{-1}$$

$$p_1 = 6 \times 10^8 \text{ nm}^{-2}$$

$$v_1 = 0.864006 \times 10^{-3} \text{ m}^3 \text{ kg}^{-1}$$

into the Rankine-Hugoniot equations for  $u$  and  $U$  give

$$u = 287.55 \text{ m sec}^{-1}$$

$$U = 2090.38 \text{ m sec}^{-1}$$

Figure 5 is a characteristic diagram for a 6 kilobar shock showing the relationships among the particle velocity ( $u_1$ ), the velocity of the compressive shock ( $U_1$ ), and the velocities of rarefaction waves associated with the shock.

The relationship between the total energy and the magnitude of an earthquake is given by the empirical equation

$$\log_{10} E = 12.24 + (1.44) (m) \quad (\text{B\aa}th, 1966)$$

where  $E$  is measured in ergs and  $m$  is the surface wave magnitude. Substitution of the magnitude 6.5 of the Chiapas shock gives an energy of  $\sim 6 \times 10^{21}$  ergs. The energy needed to heat by  $2^\circ\text{C}$  the total volume of water (diameter 60 km,

depth 10km) indicated by the ITOS-1 temperature anomaly is  $\sim 5 \times 10^{19}$  or 1% of the total energy of the earthquake.

A 6 kilobar shock propagating through the water gives a particle velocity ( $\mu_3$ ) at the air-sea interface of 288 m/sec. The speed of sound ( $a_1$ ) in air at 20°C is 344 m/sec (Handbook of Chemistry and Physics, 1959). Using shock tables (Liepmann and Roshko, 1957), it is possible to plot the particle velocity behind the shock in air as a function of shock Mach number (M) [Figure 6]. It is seen that a particle velocity of 288 m/sec corresponds to a shock Mach number of 1.65.

The entropy generated by a shock is related to the change of total pressure and is defined as

$$\frac{\Delta S}{R} = 1.65 \ln \frac{P_{01}}{P_{02}} \quad (\text{Liepmann and Roshko, 1957})$$

where  $p_0$  is the stagnation pressure and R is a characteristic gas constant. For  $M = 1.65$ ,

$$\frac{P_{02}}{P_{01}} = 0.376, \text{ giving } \frac{P_{01}}{P_{02}} = 1.14.$$

which to the first order gives  $\frac{\Delta S}{R} = 0.14$ .

The temperature change between the initial unshocked gas and the adiabatically expanded gas after the shock will be small. Therefore, it is permissible to set

$$\Delta S = \int_{T_i}^{T_f} \frac{C_p dT}{T} \sim \frac{C_p(T_f - T_i)}{T_i}$$

where  $C_p$  is the specific heat at constant pressure and  $T_f$  and  $T_i$  are the final and initial temperatures, respectively.

$$T_i \Delta S = C_p (\Delta T)$$

Taking  $\frac{T_i \Delta S}{R T_i} = 0.14$  gives

$$C_p (\Delta T) = (0.14) (R T_i)$$

$$\frac{\Delta T}{T} = \frac{0.14 (C_p - C_v)}{C_p} = 0.14 \left( 1 - \frac{1}{\gamma} \right)$$

where  $C_v$  is the specific heat at constant volume and  $\gamma$  is the ( $C_p/C_v$ ) ratio.

$$\frac{\Delta T}{T} = 0.14 \left( 1 - \frac{1}{1.4} \right) = 0.04$$

If  $T_i = 300^\circ \text{K}$ , we obtain  $\Delta T = 12^\circ \text{C}$ .

Increasing the atmospheric temperature ( $\bar{T}$ ) used in Kunde's equation by  $12^\circ$  would have the effect of raising the observed IR surface temperature to a value almost equal to the actual surface temperature ( $T_s$ ), i. e.,  $302^\circ \text{K}$ .

Truttse (1971) has suggested that increased density in the upper atmosphere is traceable to the warming of the atmosphere by shock waves generated in the troposphere during powerful earthquakes and that most spectacular effects are observed when the epicenters are located at shallow depths. Our estimates agree with Truttse's hypothesis.

#### SUMMARY

The possibility of cohesive shock waves being propagated through the ocean during an earthquake at sea was first suggested by descriptions taken from ships' logs published in Rudolphs' papers almost a century ago. Recent reports are similar (Hoffmeister, 1971). A search of infrared photographs taken by two NASA spacecraft shortly after the occurrences of some large earthquakes at sea have indicated the possibility of a thermal effect associated with the quakes. The attribution of this effect to shock waves has been explored. A consistent explanation for thermal effects of seaquakes in terms of shock wave propagation in water and air has been offered and found to be in quantitative agreement with observed results.

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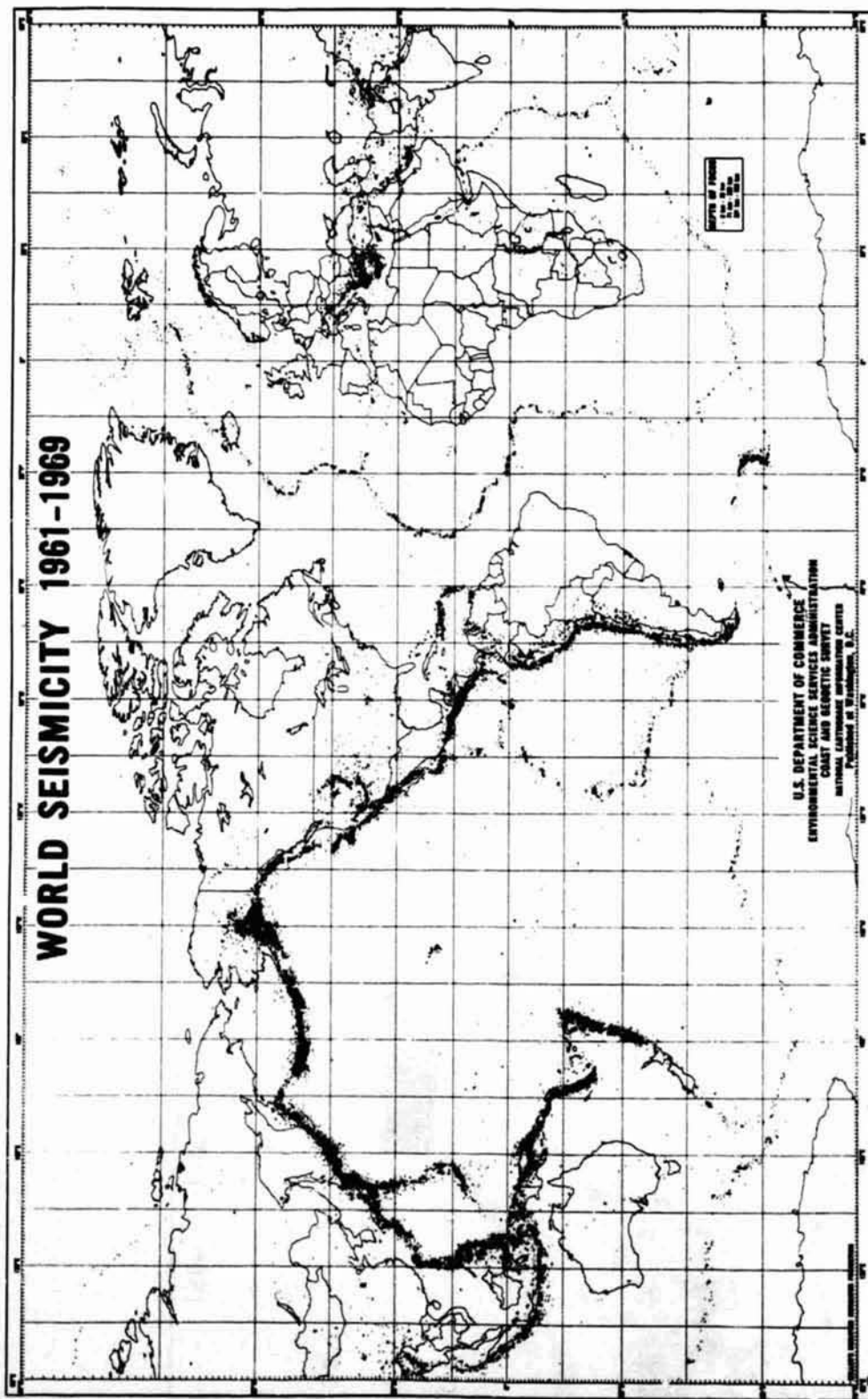


Figure 1a. Earthquake Belts

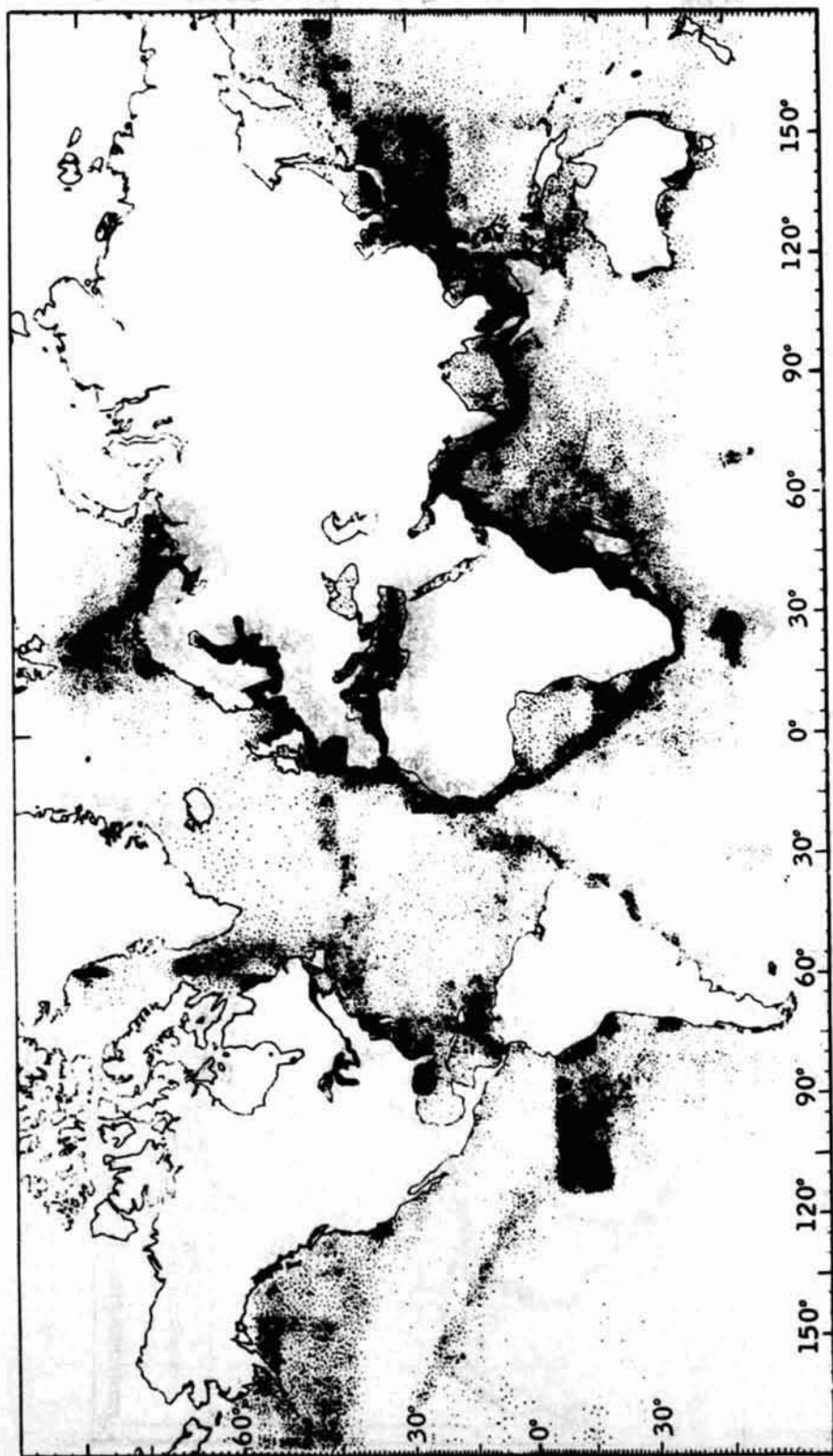


Figure 1b. Worldwide Ship Traffic Density

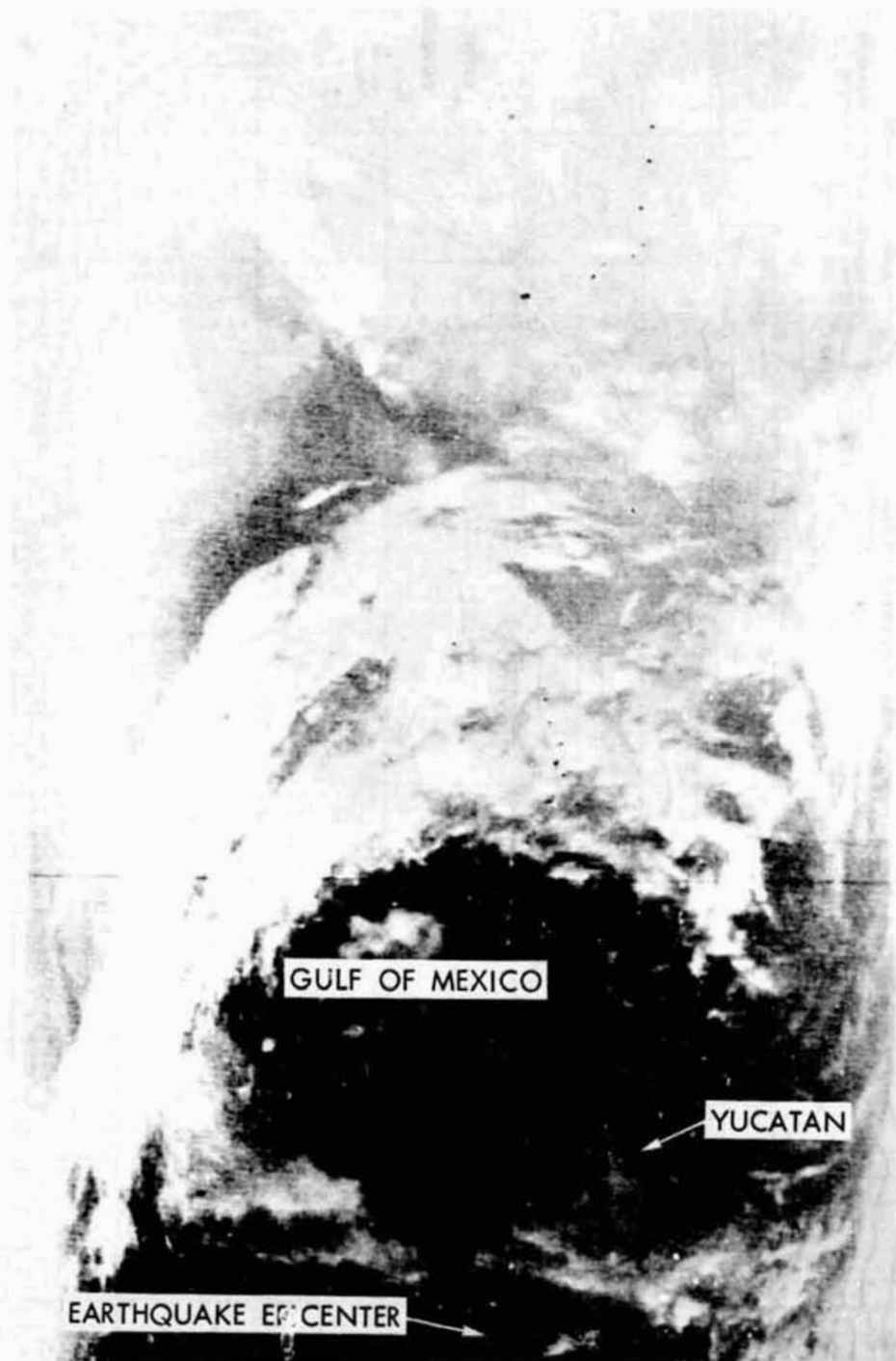


Figure 2. Pacific Ocean, Mexico, Yucatan as Seen by the ITOS-1  
11.5 $\mu$  radiometer April 29, 1970 1140 GMT



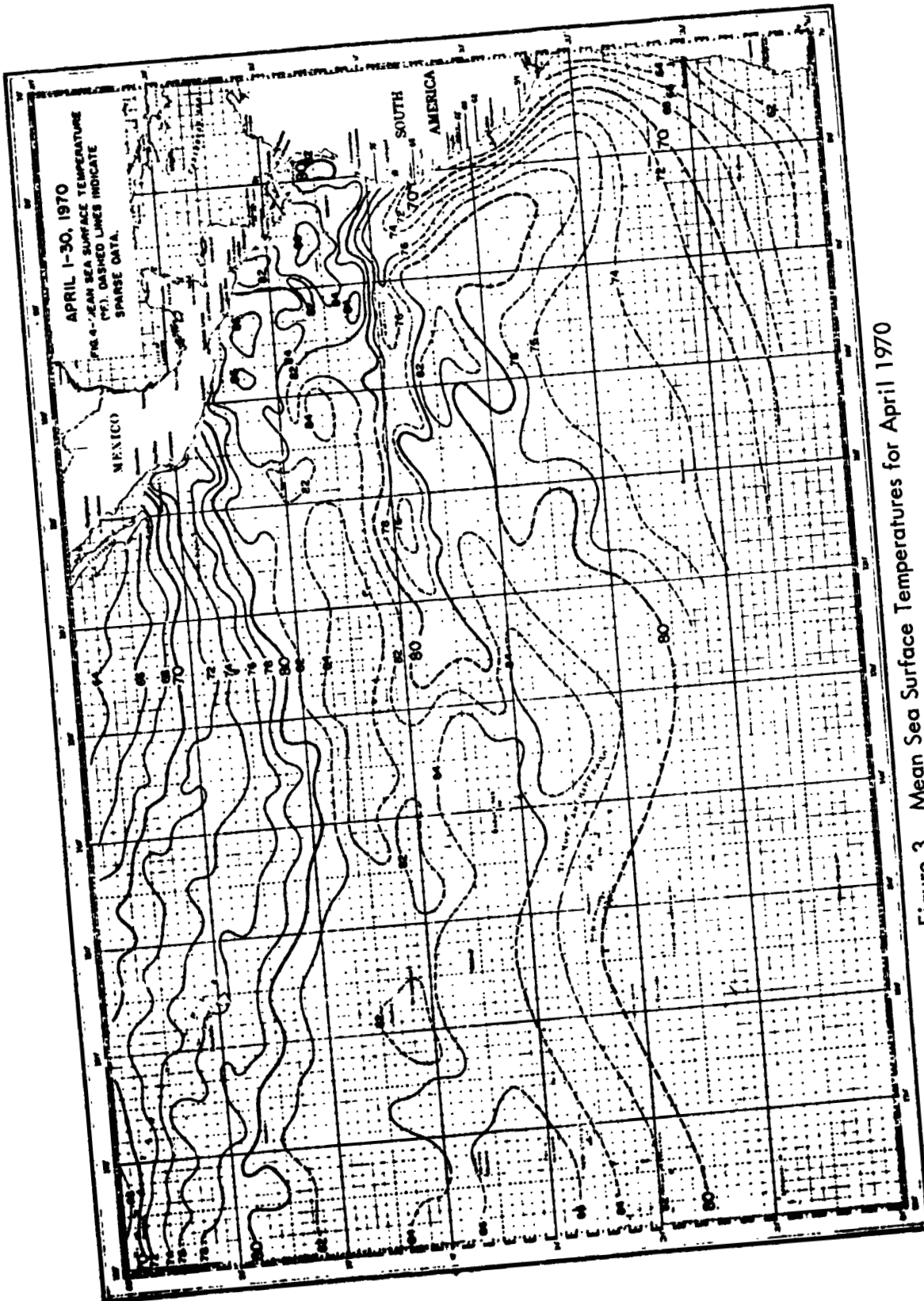


Figure 3. Mean Sea Surface Temperatures for April 1970

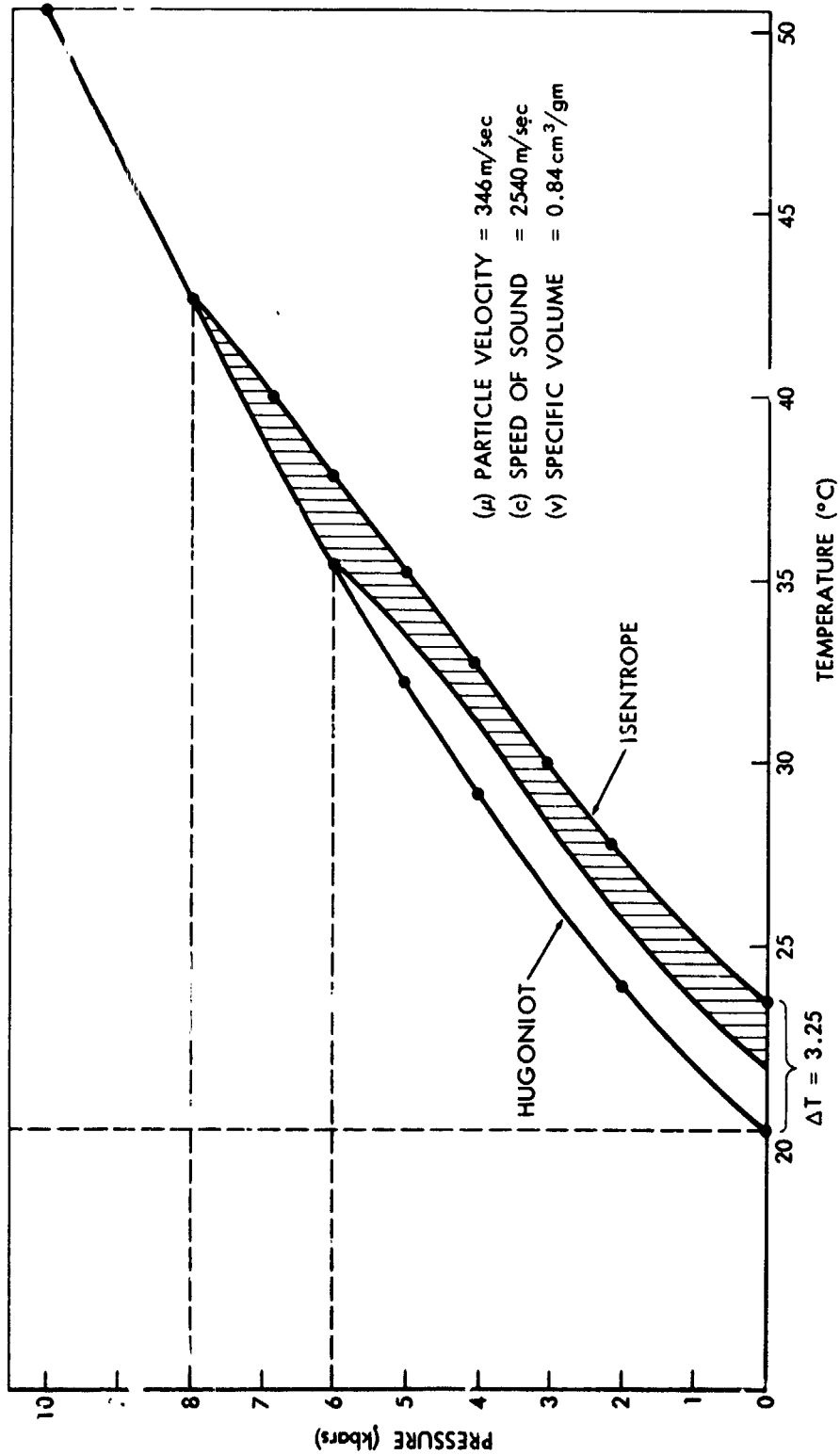


Figure 4. Hugoniot and Isentropic Curves for 6 and 8 kilobar Shocks

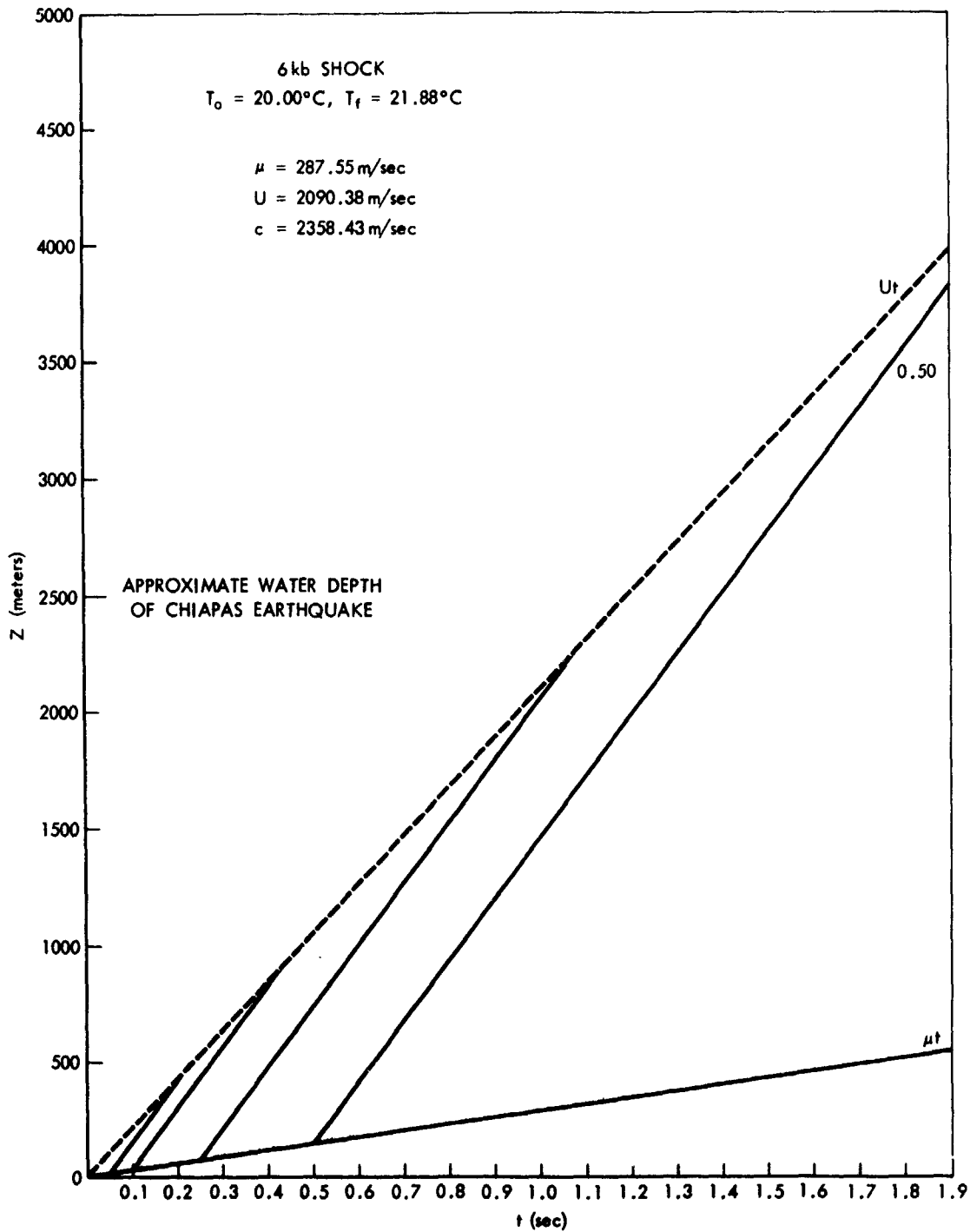


Figure 5. Characteristic Diagram for a 6 kilobar Shock Showing the Relationships among the Particle Velocity ( $\mu t$ ), the Velocity of the Compressive Shock ( $Ut$ ), and the Velocities of the Rarefaction Waves associated with the Shock.

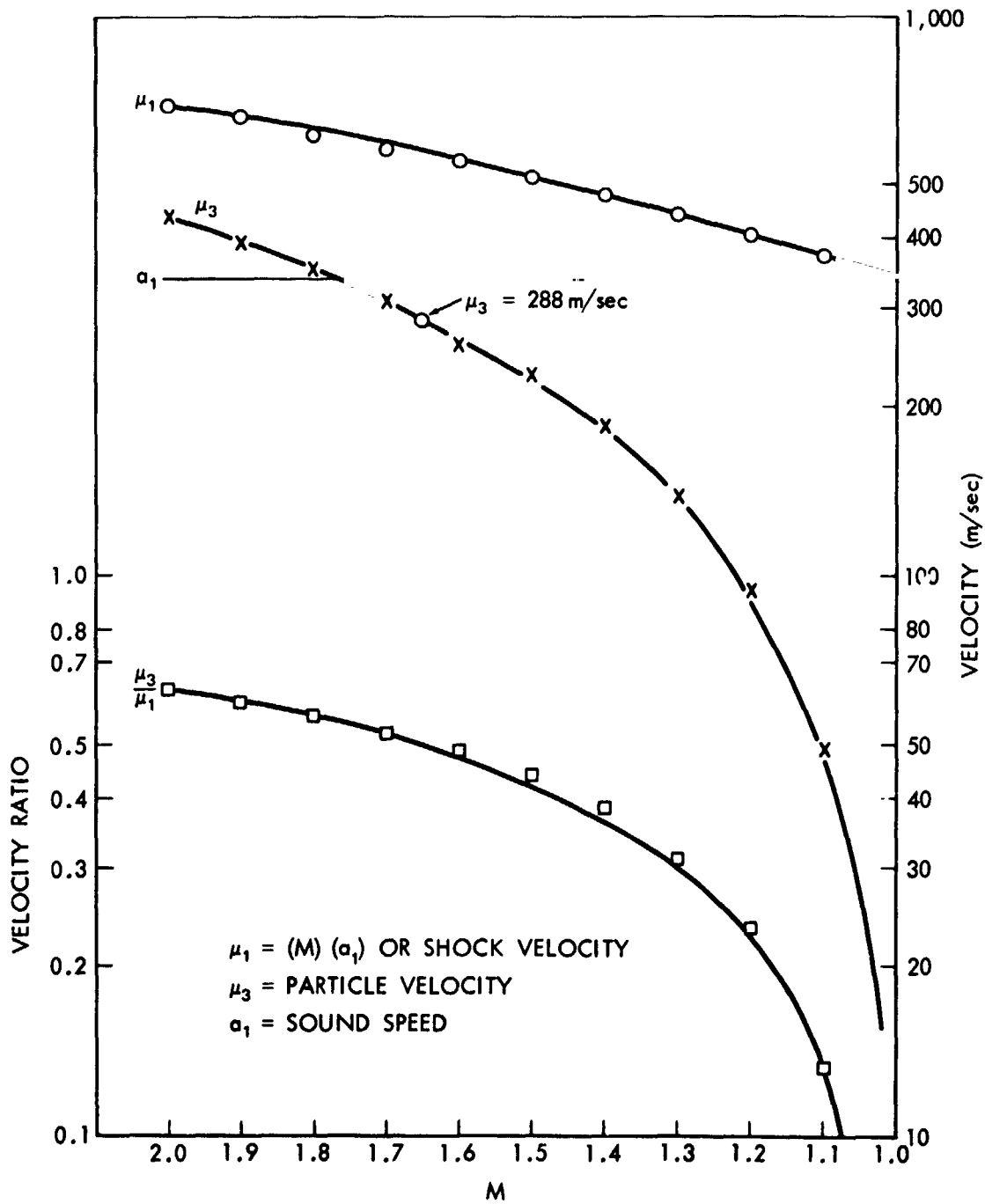


Figure 6. Particle and Shock Velocity as a Function of Mach Number