II. GENERAL INFORMATION ON EARTHQUAKES

2.1 Seismic waves

An earthquake is a sudden release of energy accumulated in rocks through the action of global tectonic processes. When the compression stresses exceed the mechanical strength of the rocks, a rupture occurs, which gives rise to seismic waves.

2.1.1 Body waves

When a disturbance occurs in a solid body, elastic waves are propagated in all directions.

* longitudinal (dilatational) waves, also called primary or P waves, in which the particles are displaced radially from the origin, i.e. in the direction of propagation;
* transverse (distortional) waves, also called secondary or S waves, in which the particles are displaced in a plane perpendicular to the direction of propagation.

Each wave travels at its own velocity (Vp for P and Vs for S), which at any given point in the medium depends on the density p and the moduli of elasticity λ and μ , and is given by the equations :

$$Vp = \sqrt{\frac{\lambda + 2\mu}{\rho}} \qquad Vs = \sqrt{\frac{\mu}{\rho}}$$

The moduli of elasticity are functions of each other and of "Poisson's ratio" σ.

$$\sigma = \lambda/2\,(\mu + \lambda)$$

In many minerals λ is approximately equal to μ and the value of σ is about 1/4; under these conditions and as a first approximation, the ratio of the velocities Vp/Vs is close to $\sqrt{3}$.

It is important to point out that the study of the variation in the ratio of velocities Vp/Vs is used for prediction purposes (see Chapter VII).

2.1.2 Surface waves

The second category of seismic waves consists of surface waves, which transmit the greater part of the energy in earthquakes having a shallow focus.

The principal types of wave are as follows :

* Rayleigh (or R) waves, which are polarized in the plane of propagation: each particle describes a retrograde eliptical orbit; the amplitude of movement is greatest at the surface and decreases exponentially in relation to the depth. In a uniform medium for which the ratio $\sigma = 1/4$, the velocity of Rayleigh waves is 0.919 Vs.
* "Love waves", which are transverse guided waves (without any vertical component).

2.2 Foreshocks and aftershocks

A major earthquake is never an isolated phenomenon. The violent and destructive main shocks may be preceded by preliminary tremors or foreshocks, which are less severe and few in number, but important to study in order to predict the destructive shocks and take protective measures (see page 42). Sometimes seismic activity in a region increases gradually in intensity up to a climax. Then there are aftershocks, which are belated shocks of decreasing intensity occurring at increasing intervals, and which suggest that the disturbed strata of the earth's crust, abruptly decompressed, recover their equilibrium only after some convulsive movements. The disturbance may last for months and keep the threatened population in a state of anxiety, which generally causes them to evacuate the area.

The pattern of foreshocks and aftershocks has been studied by Mogi, who distinguished two main types:

Type I: The strongest shock is the first, the number of aftershocks and their energy decreasing exponentially as the time (counted in days, for instance)since the main shock increases. The curve of energy released as a function of elapsed time may show some irregularities, however, as some of the aftershocks may be stronger than those immediately preceding them.

Type II: Numerous foreshocks may be observed, the number of which increases shortly before the main shock; the energy curve of the aftershocks generally decreases exponentially as in type I. Type II is often observed in the case of artificial earthquakes caused by the filling of lake-reservoirs (see page 11).

It can be seen how useful the observation of foreshocks could be for predicting a more serious shock. Unfortunately, most major earthquakes are of type I.

Within this type it is also possible to distinguish the case in which a severe shock is followed by shocks of even greater intensity in a relatively short time. The example of the Ionian Islands earthquakes is characteristic: between 9 and 31 August 1953, 338 shocks were recorded, coming from the same focus or from foci very close together; there were three climaxes <u>of increasing severity</u> at intervals of several days, the seismic energy being released in several increasingly violent shocks (9 August at 07 h 41, magnitude 6.3; 11 August at 03 h 32, magnitude 6.7; 12 August at 09 h 24, magnitude 7.1). In the present state of knowledge such phenomena are unforeseeable.

Earthquake swarms must be set apart from foreshocks and aftershocks: the energy does not decrease exponentially and the strongest shocks may occur in any order. The earthquake swarm at Matsushiro in Japan is famous: in the course of a single year 1966, 854,566 shocks were recorded at the Matsushiro observatory; 88,258 of them were felt in the area of the epicentre; only a few reached an intensity of VII.

2.3 Artificial earthquakes

For several years the attention of seismologists has been drawn to the fact that in some cases human activities, such as the filling of certain lake-reservoirs, the injection of water into deep wells or the exploitation of oil and gas deposits are followed by seismic activity. In France, for example, several earthquakes have occurred since 1969 at the Lacq oil and gas field.

The significant seismic activity accompanying the injection of waste fluid into a well sunk to a depth of 3,700 m near Denver (Colorado - USA) led scientists to study this phenomenon in the laboratory and on the ground. The injection of fluid into a faulted zone reduces the friction and thus diminishes the stresses in the fault. In simple terms, the injection of fluid weakens a fault, whereas pumping out strengthens it. Where there is a substantial stress in a fault, the injection of fluid will "release" the fault, thus causing an earthquake.

Experiments carried out at the Rangely (Colorado - USA) oilfield showed that the injection of water into wells drilled round the perimeter of the deposit caused very numerous localized shocks in the vicinity of the wells. These experiments also demonstrated the possibility of controlling seismic activity by alternate injection and pumping. These initial results suggest that it may one day be possible to control seismic activity along a major fault, such as the San Andreas fault in California.

Seismic activity observed during the filling of certain large lake-reservoirs has been the subject of increasingly numerous publications (see bibliography).

In some cases - Hsinfengkiang (China) in 1962, Kariba (Southern Rhodesia) in 1963, Kremasta (Greece) in 1966, Koyna (India) in 1967 - the magnitude of the earthquakes thus caused was over 6, and there was extensive damage.

The phenomenon seemed sufficiently serious for UNESCO to set up a working group to study it.

The weight of water may sometimes be sufficient to explain the sudden release of the stored energy. This triggering effect is favoured by the presence of strata of differing plasticity.

In addition, the rising water level in a reservoir may change the effective field of stresses in the rock mass by increasing the interstitial pressure, which sometimes causes a rupture. This phenomenon occurs, in particular, along joints, faults, and other zones of lesser strength which allow the interstitial fluid to pass. As a result of the increase in interstitial pressure, the effective normal stresses are reduced, and this may set off earthquakes; in such a case, the difference between the water level in the reservoir and the level of the natural water table is an important factor.

The recurrence of seismicity caused by the filling of a reservoir may endanger the dam itself, the population living downstream which is threatened by flooding, and neighbouring towns, where buildings may be damaged.

An on the spot survey, with an accurate evaluation of the extent of fissuring of the rocks, and the recording of tectonic irregularities (faults) as well as a closer watch on seismic activity should be obligatory in the neighbourhood of large reservoirs and should be begun well before filling.

The equipment to be used for such surveillance has been described in a document prepared by the UNESCO Working Group (SF-73/Conf.625/1).

2.4 The effects of earthquakes

The observation and study of the effects of earthquakes can provide information to serve as a basis for drawing up rules for protection. In many cases, moreover, direct observation on the ground makes it possible to determine the exact position of the epicentre with greater accuracy than can be expected from the use of methods relying solely on the interpretation of seismograms.

In any case, when the shocks are of sufficient intensity, they cause more or less extensive disturbance to the affected terrain.

2.4.1 Subsoil fissuring and fault movement

The San Francisco earthquake of 18 April 1906 rejuvenated the San Andreas fault over a length of 470 km, displacing the lips by a few centimetres up to one metre and causing a slip between the sides varying from 25 cm to 7 m. The Mino-Avari earthquake in Japan caused changes in level of about 6 m and horizontal slips of 2 m along the Neo fault which, unlike that at San Francisco, is not rectilinear, but divided into stepped sections. The consequences of such ground movements are, of course, disastrous for buildings near to or on the line of a fault. Any programme of land development should therefore carefully avoid the line of active faults when designating residential or industrial areas.

2.4.2 Lifting and subsidence

During the Tokyo earthquake (September 1923) in Japan, the coast of Sagami Bay rose 2 m, whereas on the north side of the bay the bottom rose 250 m and at the centre it subsided 100 to 200 m- even 400 m at some points. In addition, there was horizontal displacement of up to 4 m in the land surrounding Tokyo Bay. This particularly deadly earthquake killed 99,331 people, about 38,000 of whom died in the fire which swept Tokyo afterwards. This kind of destruction is much more difficult to anticipate than fault movements, and can only be mitigated, to a certain extent, by the adoption of building regulations requiring increased strength for buildings.

2.4.3 Triggering of mass earth movements

Earthquakes sometimes cause rock falls, landslides, horizontal movements of the ground comparable to the movements of materials on a vibrating table, the collapse of deeply eroded peaks, avalanches and falls of stones and seracs. Such phenomena could be detected in advance by detailed geological and geomorphological studies, which would make it possible to identify the areas likely to be devastated.

2.4.4 Liquefaction and settling

When subjected to the vibrations caused by an earthquake, certain subsoil strata consisting of loose, small grain-size materials may settle, causing subsidence of the overlying strata. Sandy formations saturated with water may exhibit liquefaction phenomena. The effect of the tremors is to increase interstitial pressure, by transferring the grain interface load to the capillary water, thus causing a sudden reduction in carrying capacity. Building in the danger areas could only be avoided by a prior survey of the loose strata, based on geophysical and geological investigation.

2.4.5 Disturbance of springs

Disturbance of the subsoil causes changes in the course of groundwater flows. This can cause abrupt changes in the level of the water table and sudden drying up of surface springs.

2.4.6 Tsunamis

The Japanese word "tsunami" designates the great waves or sequence of waves caused by disturbances of the seabed, which are often associated with earthquakes. It is not known exactly how tsunamis happen, and, as a result, they may have several sudden origins: vertical movements of the seabed caused directly by earthquakes, underwater landslides associated with an earthquake or a volcanic eruption, etc.

This phenomenon and the various methods used to combat its disastrous effects are considered at greater length in annex II (see page 113).

III. MICROSEISMIC STUDY OF EARTHQUAKES

The evaluation of the danger to the population from earthquakes depends on determining seismic areas as accurately as possible, both by the interpretation of data obtained by seismological observation (microseismology) and by direct observation on the ground (macroseismology).

The arrival time of a certain number of waves (or phases) can be accurately determined (generally to within one tenth of a second) by the interpretation of seismograms. Identification of these phases makes it possible to calculate the epicentral distance Δ (distance between the epicentre of the earthquake and the observatory) and possibly the depth of the focus.

When the earthquake occurs at a relatively short distance (less than 1,000 km) the clearest phases correspond to the Pg and Sg waves propagated in the main layer of the crust (the granitic layer) and to the Pn and Sn waves propagated below the Mohorovicic ("Moho") discontinuity, which separates the crust from the "upper mantle" of the earth.

As the velocities of the P and S (Pg and Sg, Pn and Sn) waves differ, the difference between the arrival times of the longitudinal and transverse waves will increase as the epicentral distance increases; reading this time difference on a seismogram thus makes it possible to find the epicentral distance immediately, by reference to tables drawn up for particular models (Jeffreys, Haslach, Balkans, etc.).

When the earthquake occurs at a distance of more than 1,000 km, the P and S phases will be used, the P wave being particularly clear on seismograms produced by instruments recording the vertical component of the ground movement, whereas the S wave is more clearly visible on recordings of the horizontal component.

Several tables of propagation-times are available, such as the Jeffreys-Bullen tables (published by the International Seismological Summary in 1940) or the more recent tables of Herrin, which have been improved, in particular, thanks to the data furnished by large chemical and nuclear explosions which constitute surface sources accurately located in space and time. These tables give the average propagation-times of the principal phases as a function of the angular distance in degrees and the focal depth.

3.1 Determination of the parameters of an earthquake focus: geographical co-ordinates of the epicentre, focal depth

3.1.1 Case of a single station

The epicentral region can be roughly located from the recordings of a single station: the epicentral distance is determined from time-differences such as S-P, and the azimuth by the ratio of the amplitudes A_E and A_N of the first impulse on the recordings of movement components in the directions EW and NS, taking account of the direction of the vertical component of this first movement.

The azimuth can be more accurately determined by setting up tripartite stations or complex stations (arrays). A tripartite station consists of three observation posts provided with short-period vertical seismographs of high sensitivity (magnification of up to 250,000 to 500,000), placed at the apex of a triangle, at distances of 30 to 50 km.

The three stations are linked by teletransmission and the three recordings are made on the same diagram at the principal station. From the differences in the arrival-times of the same wave at the three instruments, the azimuth of the source and the apparent speed of propagation of the wave can be determined: the distance of the source can be calculated approximately from the figure found for this speed.

The complex stations (arrays) installed in some countries comprise about a hundred seismographs distributed over the ground inside a circle which may have a radius of up to about 50 kilometres. Combination of the recordings gives greater phase clarity (impetus) against the background noise. Despite these improvements, experimental results show that the calculated positions of epicentres may sometimes be several hundred kilometres distant from the true position.

3.1.2 Case of several stations (earthquakes nearby)

The parameters of an earthquake nearby can be determined geographically from the data supplied by a regional network (hyperbola method, using the differences in arrival-times of the Pg phase at pairs of stations, $S_1 - S_2$, $S_2 - S_3$, etc.) or by computer. In the latter case, a regional table of propagation-times is introduced into the general programme for determining epicentres.

One such programme has been published, in a report prepared for UNESCO (International Seismological Centre, Epicentre Determination Programme and Handbook, UNDP-UNESCO project: Study of seismicity in the Balkan region, Strasbourg - France, 1 October 1972).

3.1.3 Case of several stations (distant earthquakes)

Several national or international services regularly calculate focus parameters; this is now done by computer:

1) The International Central Office of Seismology, 5 rue René Descartes 67000 Strasbourg (France) circulates epicentre determinations for Europe, the Mediterranean basin and Western Asia within a few days.

2) The U.S. Geological Survey, National Earthquake Information Service D2, Denver Federal Centre, Blgd. 25, Denver, CO 80225, USA, publishes within a few weeks, many preliminary determinations of epicentres(in all about 5,000 a year) for all parts of the world.

3) The International Seismological Centre, 6 South Oswald Road, Edinburgh (United Kingdom), publishes monthly and within a few months of the event, a large number of final determinations of epicentres, also for all parts of the world.

3.1.4 Depth of Foci

The precise determination of the depth of an earthquake focus is a problem still not completely solved.

In the case of "intermediate" earthquake (depth between 70 and 300 km) and deep earthquakes (depth between 300 and 725 km) the depth can be estimated by direct reading of recordings. This method uses the difference in the arrival-times of the P wave and the pP wave (wave reflected near the epicentre and characteristic of intermediate and deep earthquakes): for the same epicentral distance, this time-difference increases rapidly with the depth of the focus. The depth can be determined in this way to within about 10 kilometres.

Computer programming provides for determination of the depth at the same time as the other parameters (longitude, latitude, time, origin); this calculation is only possible if the stations providing the data are placed at regularly increasing distances from stations located as close as possible to the epicentre. Possible errors may sometimes reach several tens of kilometres.

When the focus of an earthquake is in the earth's crust, for a known epicentral distance, the difference in the arrival-times of the Pn and Pg waves at the same station may be used. This time-difference, as shown in the tables of propagation-times, naturally depends on the model chosen and, in particular, on the depth of the Mohorovicic surface. Thus the result is only approximate and large errors are possible.

It should be realized how uncertain the determination of focal depth still is at present. In view of the value of such determination for the tectonic interpretation of seismicity, it is desirable that further research should be undertaken on the subject.

3.2 Magnitude

Magnitude is a quantity representing the energy generated at the focus of an earthquake. This concept was first proposed by C.F. Richter in 1934, on the basis of the following observation.

Suppose there are several geophysical stations, each equipped with a seismograph (the standard instrument is the Wood-Anderson torsion seismometer, capable of recording the maximum horizontal movement of the ground A).

When there is an earthquake, the epicentre is first localized by appropriate analysis of the recorded values of A. A graph is then plotted, point by point, taking the distance epicentre to station as abscissa and the movement A as ordinate (fig. 3-1).

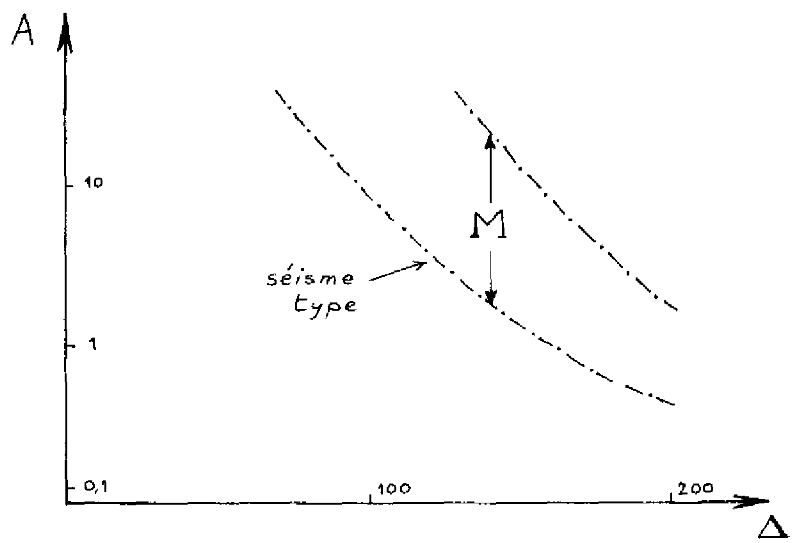

Fig. 3-1.

It has been found experimentally that the curves thus obtained are practically parallel to each other, so that if the graphs of two different earthquakes are compared, the difference between them will be found to be the same whatever the abscissa.

On the basis of this observation, a standard earthquake-or rather a standard diagram - has been established. Beside this is drawn the empirical diagram obtained after an earthquake from the observations recorded at the seismographic stations. The distance M between the two curves is taken as the measure of the energy of the earthquake. The term"magnitude" has been proposed for this value.

The concept of local intensity I, measured on the Mercalli scale, will be introduced later. It may be noted at once that a direct correlation between the two values I and M is subject to many uncertainties. For instance, an earthquake of surface origin with effects limited to a radius of a few km (as at Tuscania, Italy, in 1970) may have destructive effects at the epicentre although its total energy is low (M=3.5 at Tuscania). On the other hand, an earthquake with a deep focus - 70 to 100 km below the ground surface - may have only slight destructive effects, spread over a large area. In this case, the Mercalli intensity at the epicentre is low (e.g. I = VIII), but the magnitude M may be as much as 5 or 5.5.

An earthquake of magnitude 3 corresponds to a shock felt over a small area; an earthquake of magnitude 4.5 may cause slight damage; one of magnitude 6 - that of the Skopje earthquake in 1963 - will cause extensive damage, especially if the focus is not deep; the San Fernando earthquake (California, 1971) which had the relatively small magnitude of 6.6, but struck a highly industrialized area, caused damage estimated at $ 550 million. The greatest earthquakes, recorded by all stations of the world, have magnitude between 7 and 8.6; it is thought that the earthquake having the greatest magnitude of all, nearly 9, was at Lisbon in 1755.

The magnitude M is related to the energy released at the focus of the earthquake by the following approximate formula :

$$\log E \text{ (ergs)} = 11.8 + 1.5 M)$$

This formula makes it possible to calculate the seismic energy released in each region during a given time interval; it shows that an earthquake of magnitude 8.5 is 100 million times stronger than a small earthquake of magnitude 3.

A rough comparison may be made with the energy developed in nuclear explosions: the magnitude of an A bomb (equivalent to 20.000 tons of TNT) is about 6; a 50 megaton H bomb would have a magnitude of over 8, that is to say, similar to that of the greatest earthquakes.

The exact determination of magnitude from amplitudes measured on recordings is often difficult, and it is not surprising that differences of up to one or even two units may be found in the magnitude values for the same earthquake calculated at different stations; the effect of the subsoil at the station and that of the nature of the wave-path (sub-oceanic or continental) between the focus and the station explain these differences. It is preferable to rely on the magnitude calculated at the Pasadena base station.

IV. MACROSEISMIC STUDY OF EARTHQUAKES

Earthquakes are studied by observation and then by analysis of their effects on the ground, on buildings and on man; the effects considered here are those directly observable without the use of any special **instruments.**

For "historic" earthquakes (those before the beginning of the twentieth century, when observation with instruments began), the description of their effects is the only evidence we have for determining the epicentre and approximate magnitude, and for defining the seismicity of different regions.

Earthquakes perceptible to man are designated macroseismic. The area over which direct human observations can be made varies widely according to the energy developed at the focus of the earthquake and according to the focal depth. Some earthquakes are felt in only a few villages, whereas others are felt over large areas: the Assam earthquake (15 August 1950), 3,000,000 km2; the Kansu earthquake in China (16 December 1920), 4,000,000 km2.

The limits of the area of perception of an earthquake are often very difficult to define exactly; they depend on the individual acuity of the senses of different observers. As with many geophysical problems, it is necessary to take averages and allow for a number of observations.

Macroseismic investigation depends on evaluation of the intensity of an earthquake at a given point. It is knowledge of the average intensity of earthquake in each region that will be used to measure the danger to which this phenomenon exposes man and his works. Efforts have therefore been made progressively to establish a scale of intensity which is accessible to everyone, applicable everywhere, and which enables an observer without equipment to indicate easily the intensity of an earthquake at the point of observation.

4.1 Intensity scales

Many scales of intensity have been used; among them are the Rossi-Forel scale (RF) of 10 degrees, which is still used in several countries, and the Mercalli modified scale (MM) of 12 degrees, which is used, in particular, in the United States of America. The most recent is the MSK scale, proposed in 1964 by Medvedev, Sponheuer and Karnik.

At the lower degrees of intensity, the scales differ very little:

- I: Shock not felt;
- II: Shock felt by a few people, especially on the upper floors of buildings;
- III: Shock felt by some people indoors, vibration of windows and swaying of objects;
- IV: Shock felt by many people indoors, creaking of floors and walls, vibration of windows, doors and crockery;
- V: Shock felt by the whole population of a locality, many sleepers awakened, spilling of liquids, wide swinging of hanging objects.

At degrees VI to X, the MSK scale adds precision to the definitions of the other intensity scales by taking account of types of construction, the percentage of buildings damaged and the nature of the damage.

4.1.1 Types of construction

Type A: cobwork, adobe, rural and ordinary stone buildings;

Type B: brick, concrete block, combined masonry and wood, bonded stonework;

Type C: reinforced concrete and strong wooden buildings.

4.1.2 Percentage of buildings damaged

Q (a few) : about 5 per cent;

N (many) : about 50 per cent;

P (most) : 75 per cent or more.

4.1.3 Nature of damage

1: cracking, fall of debris and plaster;

2: cracking of walls, fall of roof tiles, cracking and fall of parts of chimneys;

3: deep and wide cracks in walls, fall of chimneys;

4: breaches in walls, partial collapse of buildings, destruction of backing or internal walls;

5: total collapse of buildings.

The following table makes it possible to determine accurately the intensity of an earthquake at a given point, according to the type of construction (A,B,C), the percentage of buildings damaged (Q,N,P) and the nature of the damage (1 to 5) :

Intensity	Type A Q	N	P	Type B Q	N	P	Type C Q	N	P
VI	2	1		1					
VII	4	3		2				1	
VIII	5	4		4	3		3	2	
IX		5		5	4		4	3	
X			5		5		5	4	

The effects on the ground increase with the intensity of the earthquake. Thus, at intensity VI, small fissures appear in wet ground; at intensities VII and VIII, the flow of springs changes, roads are cracked and the water of lakes is muddied by stirred-up silt; at intensity IX, water, sand and mud are thrown about, rocks fall and there are many landslides; at intensity X, bridges, dams and dykes are damaged, underground pipelines are broken, railway lines are twisted and fissures may be up to 1 m wide.

At intensity XI, even well-constructed buildings are severely damaged; serious ground deformation occurs and there are many landslides and rockfalls. At intensity XII, the whole landscape changes: the topography is disrupted and enormous fissures are formed; valleys blocked by landslides become lakes; structures above and below ground are seriously damaged or destroyed.

4.2 Isoseismal maps

A macroseismic study is carried out either by direct investigation on the ground, or by sending out questionnaires to authorities (mayors, teachers) in the affected area. Once the intensities observed at different points have been marked on a map, it is possible to draw isoseismal lines, which delimit areas in which the same intensity has been observed (fig. 4-1). The macroseismic epicentre lies within the pleistoseismal area bounded by the isoseismal of highest intensity. The shape of the isoseismal lines provides information on the influence of the terrain on the propagation of seismic movement. The nature of the subsoil plays an important part: intensity is usually greater in loose, alluvial ground. If the isoseismals are close together and elongated, they suggest the existence of a tectonic irregularity (fault), which is the site of the earthquake. The shape and spacing of the isoseismals depend on the depth of the focus. Formulae have been devised by which this depth can be calculated approximately.

One of them is Gutenberg's formula:

$$\frac{r}{h} = \sqrt{10^{\left(\frac{10}{3} - \frac{1}{2}\right)} - 1}$$

where h is the depth of the focus, r is the radius of the macroseismic area and I_0 is the maximum intensity at the epicentre.

4.3 Intensity-acceleration and intensity-velocity relations

Although it is questionable whether macroseismic data should be compared with instrumental data, it is permissible for practical reasons to try to relate intensity to maximum acceleration or ground velocity. The following table gives values for the acceleration γ of earth motion (for periods between 0.1 and 0.5 seconds) and velocity v (for periods between 0.5 and 25 seconds) at different degrees of intensity:

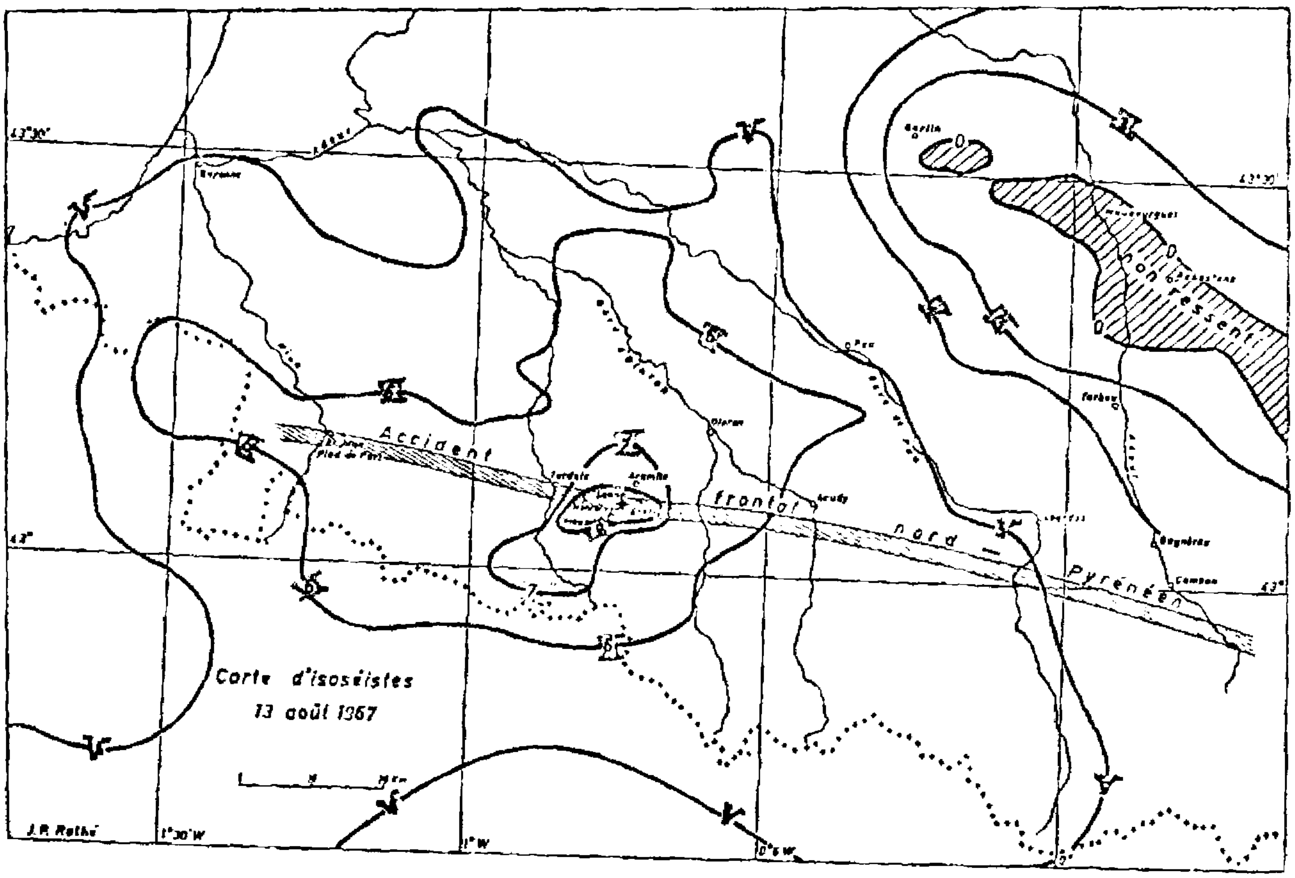

Fig. 4-1 : Example of an isoseismal map: the Arette (France) earthquake of 13 August 1967. The epicentre is located on the northern Pyrenees frontal fault. (J.P. Rothé, 1972)

I (degrees)	γ (cm/sec^2)	v (cm/sec)
V	-	1 - 2
VI	-	2.1 4.0
VII	-	4.1 8.0
VIII	-	8.1 - 16
IX	-	16.1 - 32
X	-	32.1 - 64

Reference to fig.4-1 shows that there can be different effects on the same isoseismal, depending on the different types of sub-soil. The sketch (fig. 4-2) shows the dispersion that may be observed between these parameters, even at sites in close proximity. This point will be taken up again in the section on "microzoning", which describes techniques for assessing the effects of the subsoil on surface acceleration.

Mention may also be made of the simple formula proposed by Gutenberg for California :

$$\log \gamma_o = \frac{I_o}{3} - \frac{1}{2}$$

which relates maximum acceleration γ_o to maximum intensity I_o.

4.4 Intensity-magnitude relations

In order to make use of historical observations, an attempt has been made to establish an approximate relation between magnitude M and intensity I_o:

$$M = 1 + \frac{2}{3} I_o$$

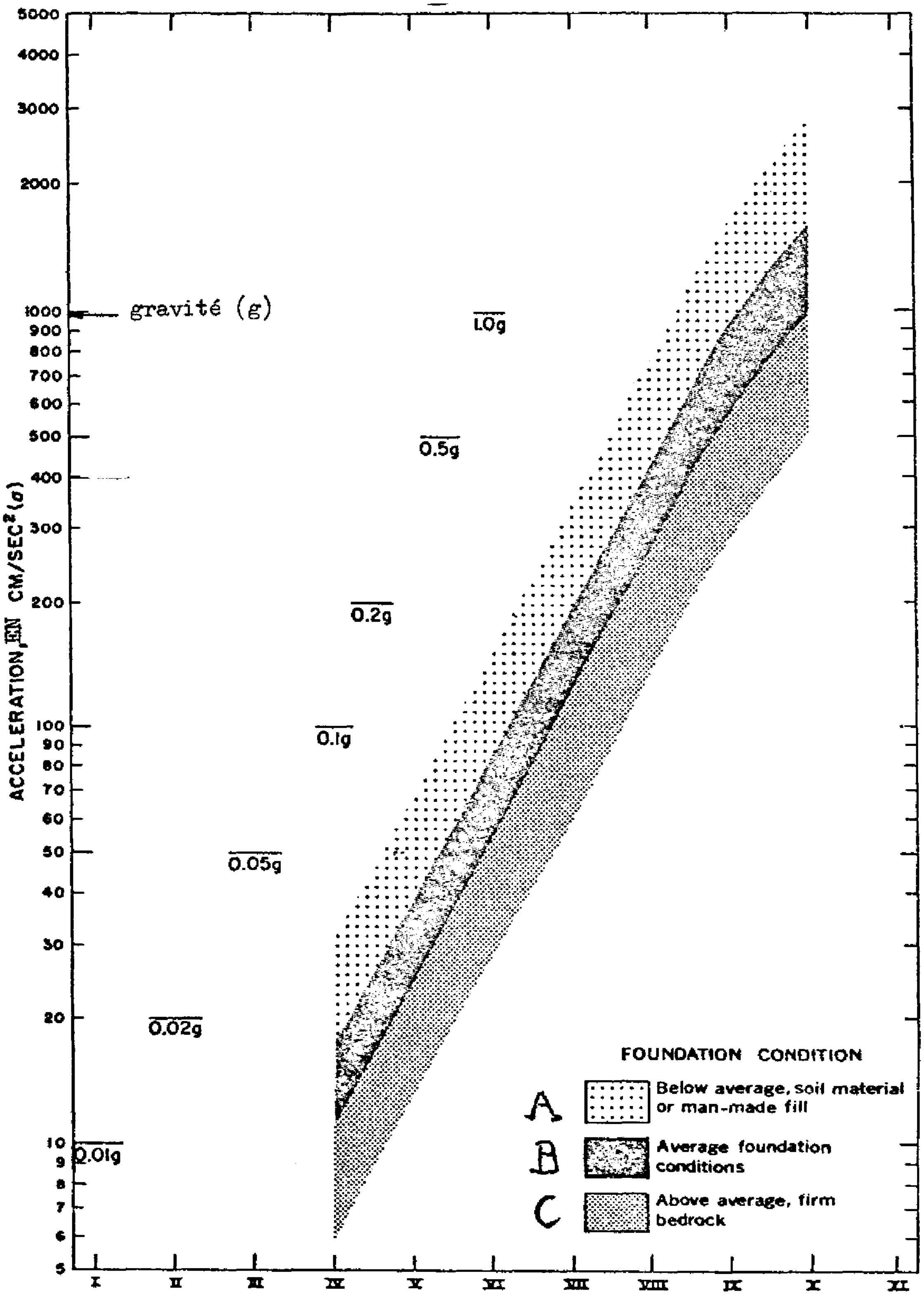

Fig. 4-2 : Relation between acceleration and macroseismic intensity (Coulter and al., 1973).

V. MEASUREMENT OF ACCELERATION - RESPONSE SPECTRUM

Earthquake engineers and seismologists need data on the intensive earth movements caused by earthquakes, particularly in the epicentral zone. Direct measurement of the acceleration of earth motion and the study of its frequency spectrum are essential elements for inclusion in engineers' calculations.

Several types of instruments (strong-motion seismographs) are now in use. These are generally accelerographs with a single degree of freedom which record the two horizontal components and the vertical component of the motion. The maximum accelerations recorded may range from 0.5 g to 2 g; the range of usable frequencies may extend from about 0.05 to 30 Hz (which corresponds to periods of ground movement varying from 0.03 to 20 seconds).

The problems raised by the measurement of high-intensity earth motion were discussed at a symposium held at Mexico City in August 1972 under the auspices of UNESCO (see Strong Earthquake Motion, Nature and Resources, vol. IX, No. 4, UNESCO, Paris, 1973, pp. 12-16).

Recordings made during recent earthquakes show that the accelerations measured were greater than had generally been assumed. An acceleration of 0.4 g was recorded in the epicentral area of the Managua earthquake (23 December 1972), an acceleration of 0.5 g on the Koyna dam (10 December 1967) and even a value of 1 g on the Pacoima dam (the San Fernando, California, earthquake of 9 February 1971). These results are particularly important for earthquake engineering calculations and it is desirable that the number of instruments in service be considerably increased so that sufficiently abundant information may be made available to engineers.