



Raggi sismici attraverso la Terra sferica

Poichè la velocità delle onde sismiche in generale aumenta con la profondità, I raggi sismica saranno concave verso l'alto (vedi parametro del raggio). I vari raggi sismica sono classificati come segue:

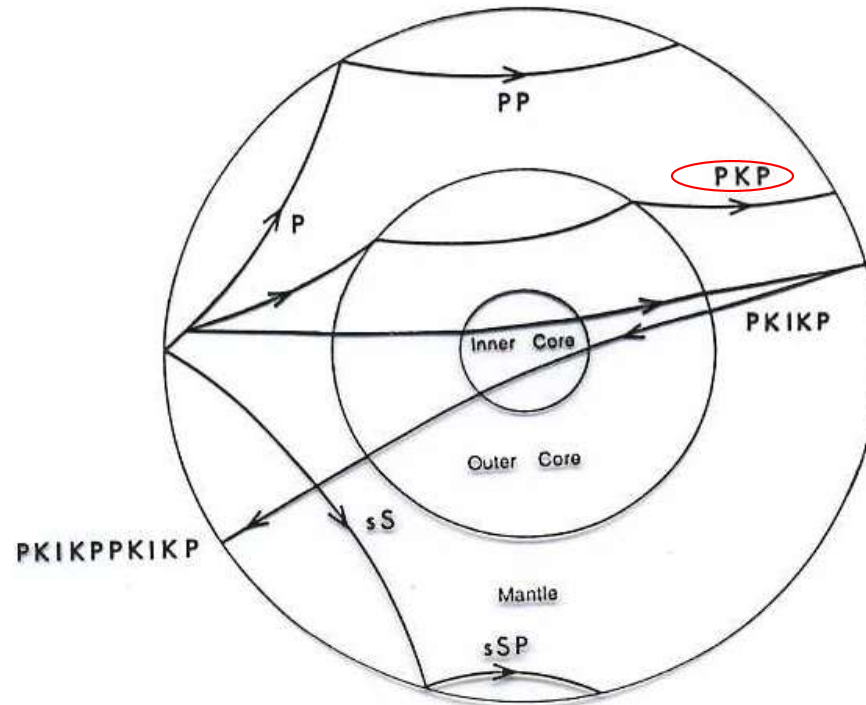
P	onda P nel mantello
S	onda S nel mantello
K	onda P nel nucleo esterno
I	onda P nel nucleo interno
r	riflessione della discontinuità mantello – nucleo esterno
i	riflessione della discontinuità nucleo esterno – nucleo interno
p	onda P riflessa dalla superficie terrestre vicino alla sorgente
s	onda S riflessa dalla superficie terrestre vicino alla sorgente
LR	onda superficiale di Rayleigh
LQ	onda superficiale di Love



Raggi sismici attraverso la Terra sferica

Esempi

L'onda **PKP** è un'onda **P** che si è propagata verso il basso nel mantello, è entrata nel nucleo esterno ed è risalita attraverso il mantello;

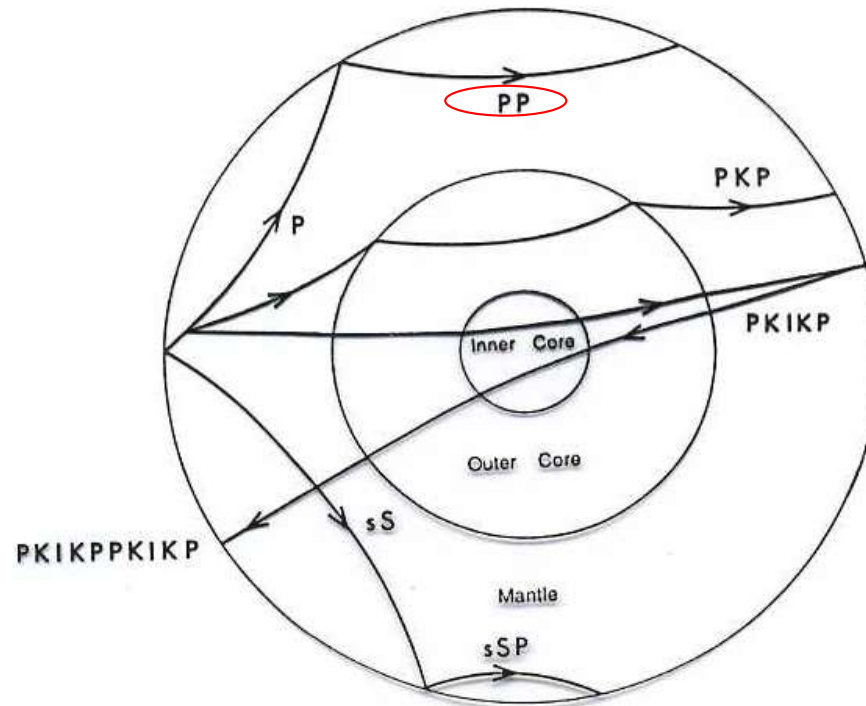




Raggi sismici attraverso la Terra sferica

Esempi

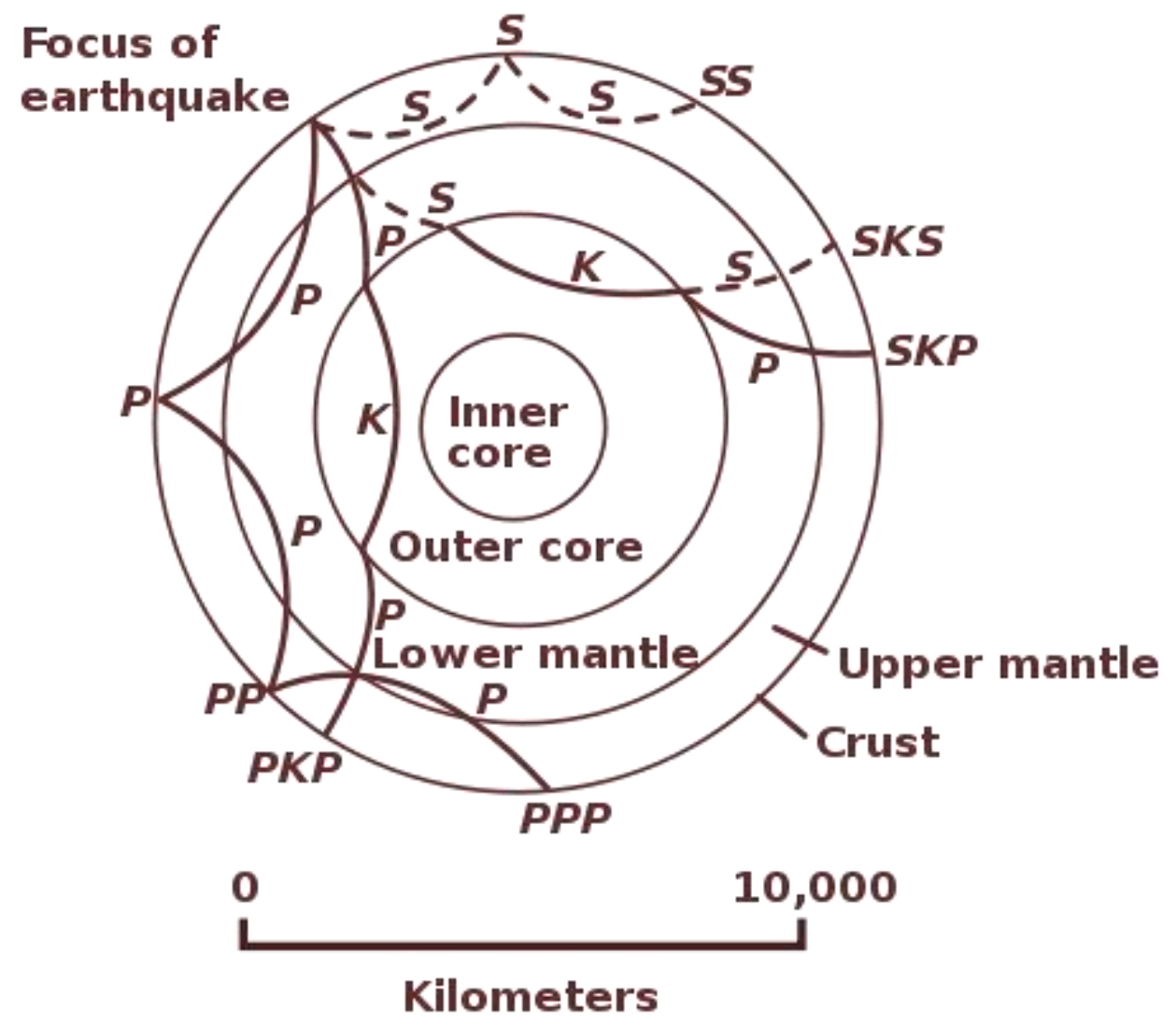
L'onda **PP** è un'onda **P** che si è propagata verso nel mantello, e che è stata riflessa una volta dalla superficie terrestre;





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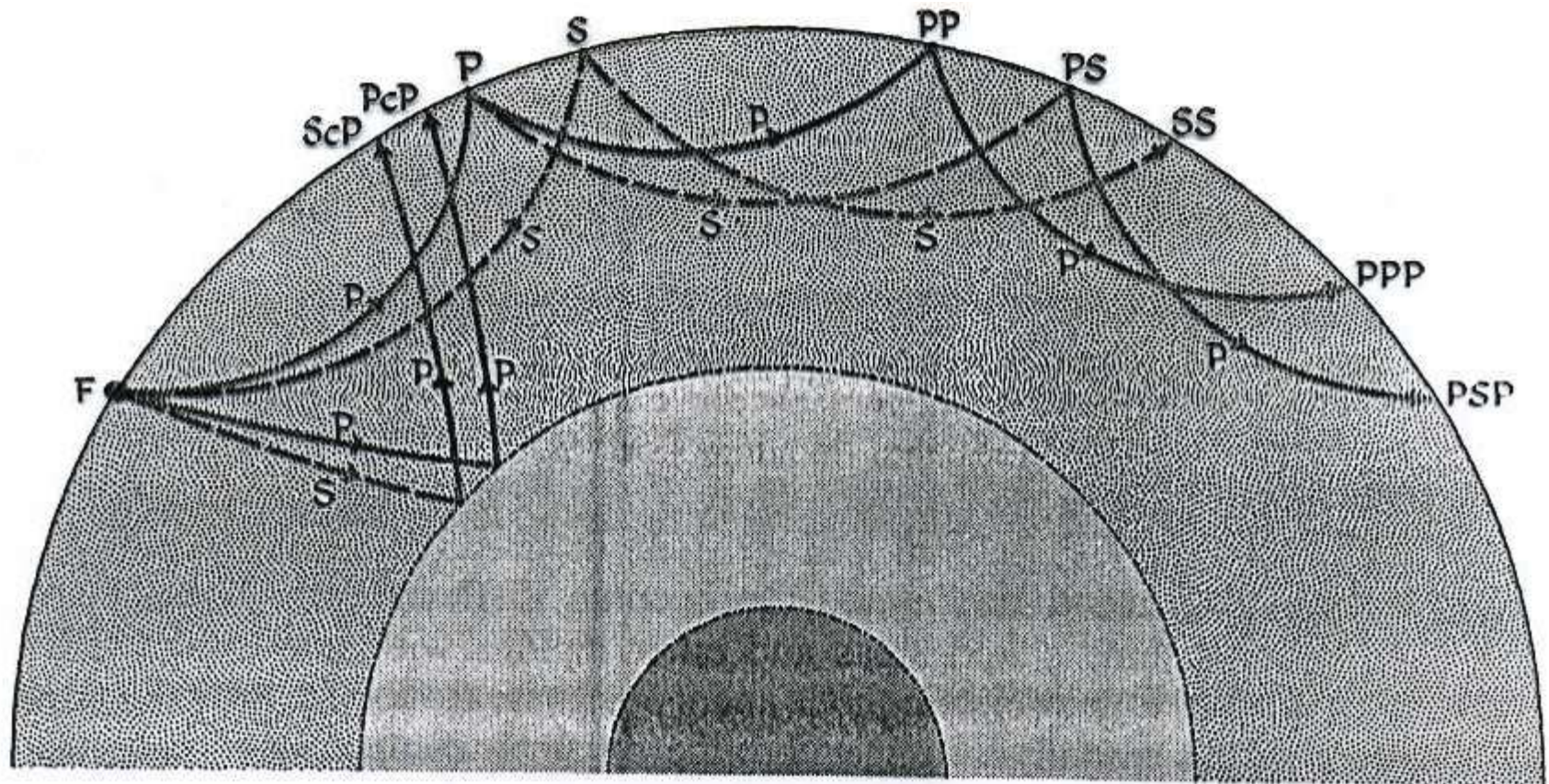


Fig. 14. Examples of propagation paths of direct and reflected waves in the Earth's mantle. Solid and dashed rays are used to distinguish between P and S waves, respectively. Waves are generated by the surface focus, F , of the earthquake which radiates both P and S waves. Different shadings show the mantle, outer core and inner core. For notation see the text.

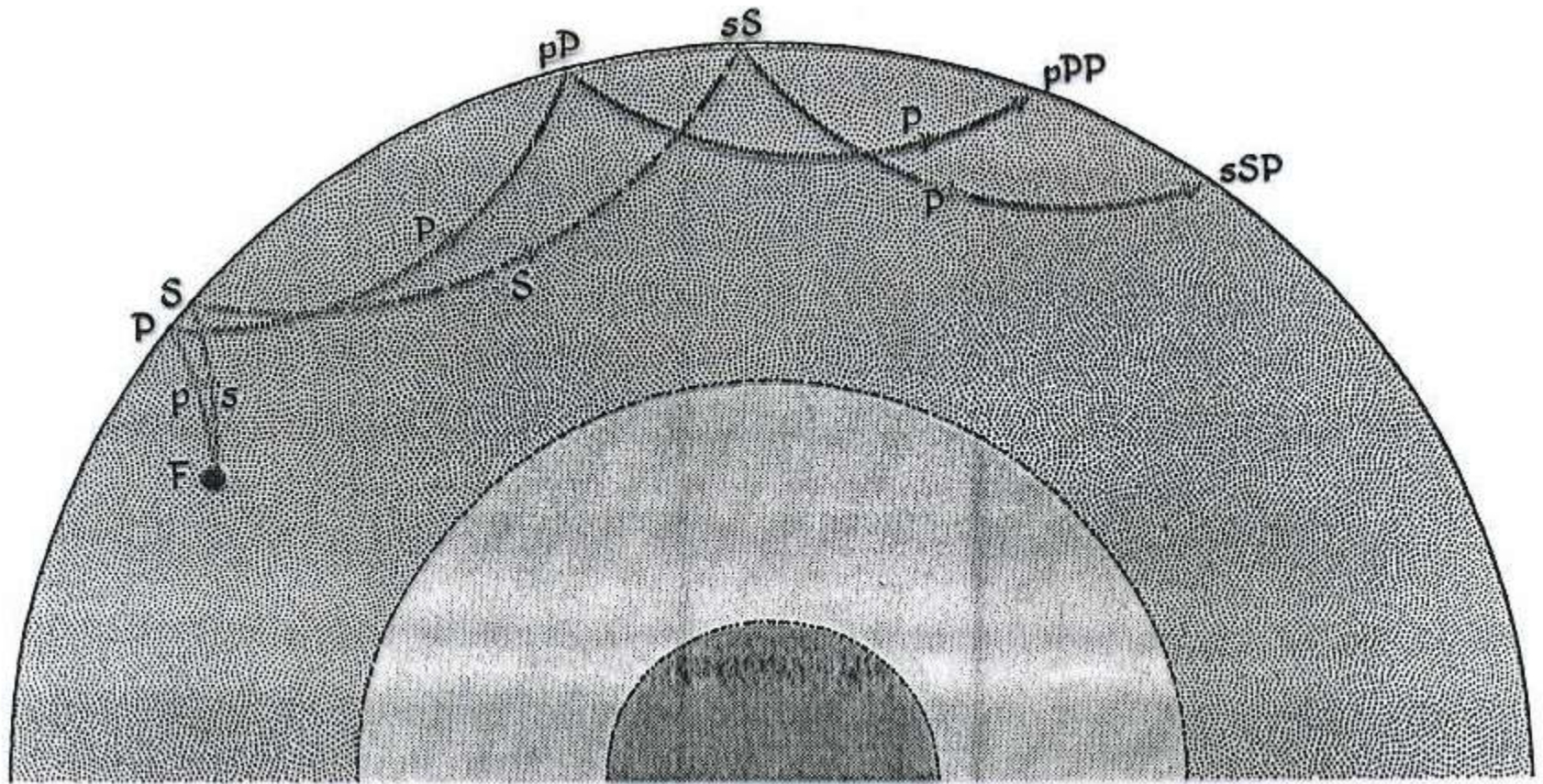


Fig. 15. Examples of propagation paths of depth phases and their notation. Waves begin at the deep focus, F , of the earthquake. Conventions as for Fig. 14.



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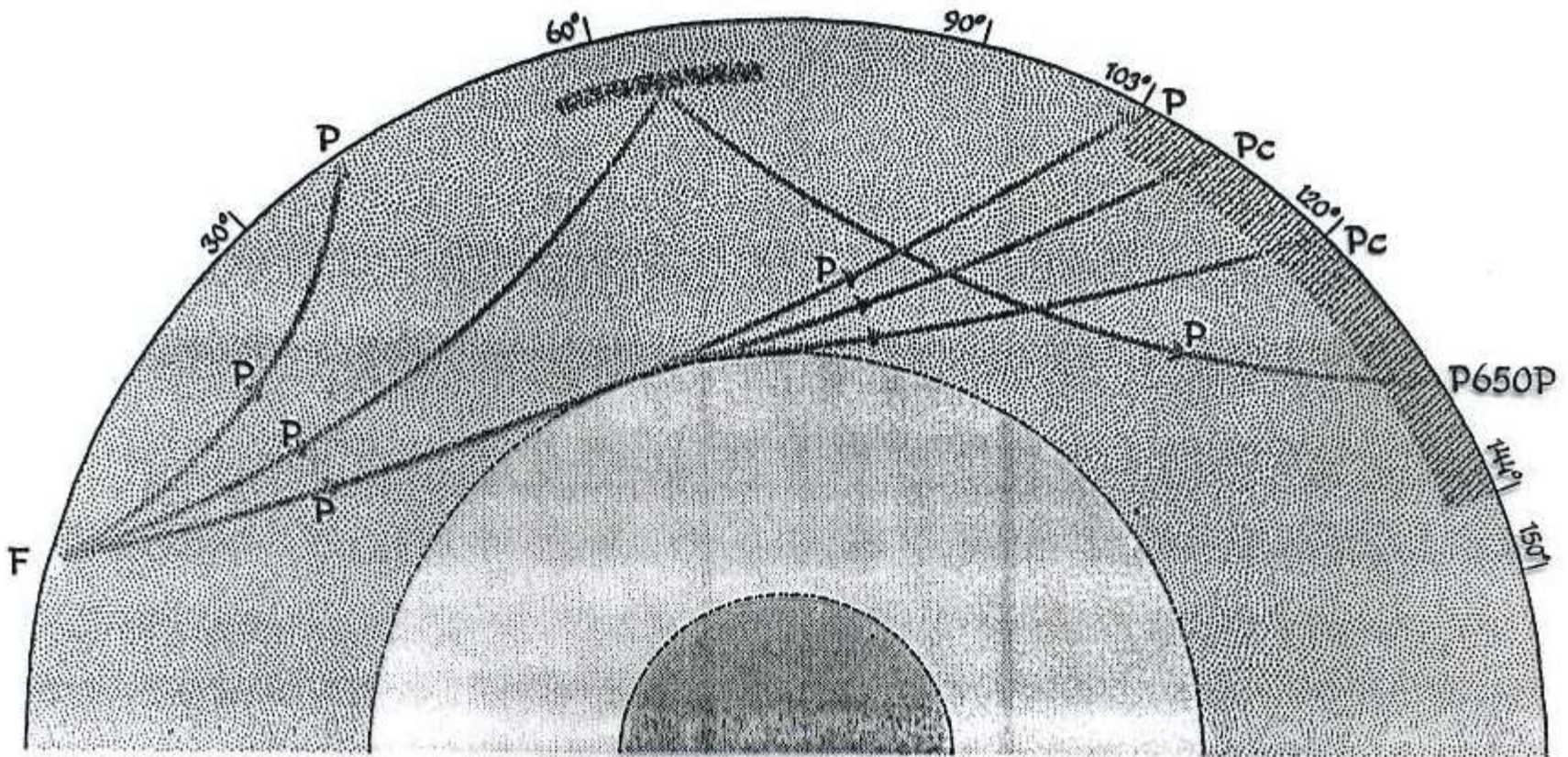


Fig. 16. Examples of propagation paths of direct P waves, P waves diffracted around the core-mantle boundary and P waves reflected downwards at a discontinuity at 650 km depth. The discontinuity and the shadow zone (103-144°) are shaded. Conventions as for Fig. 14.



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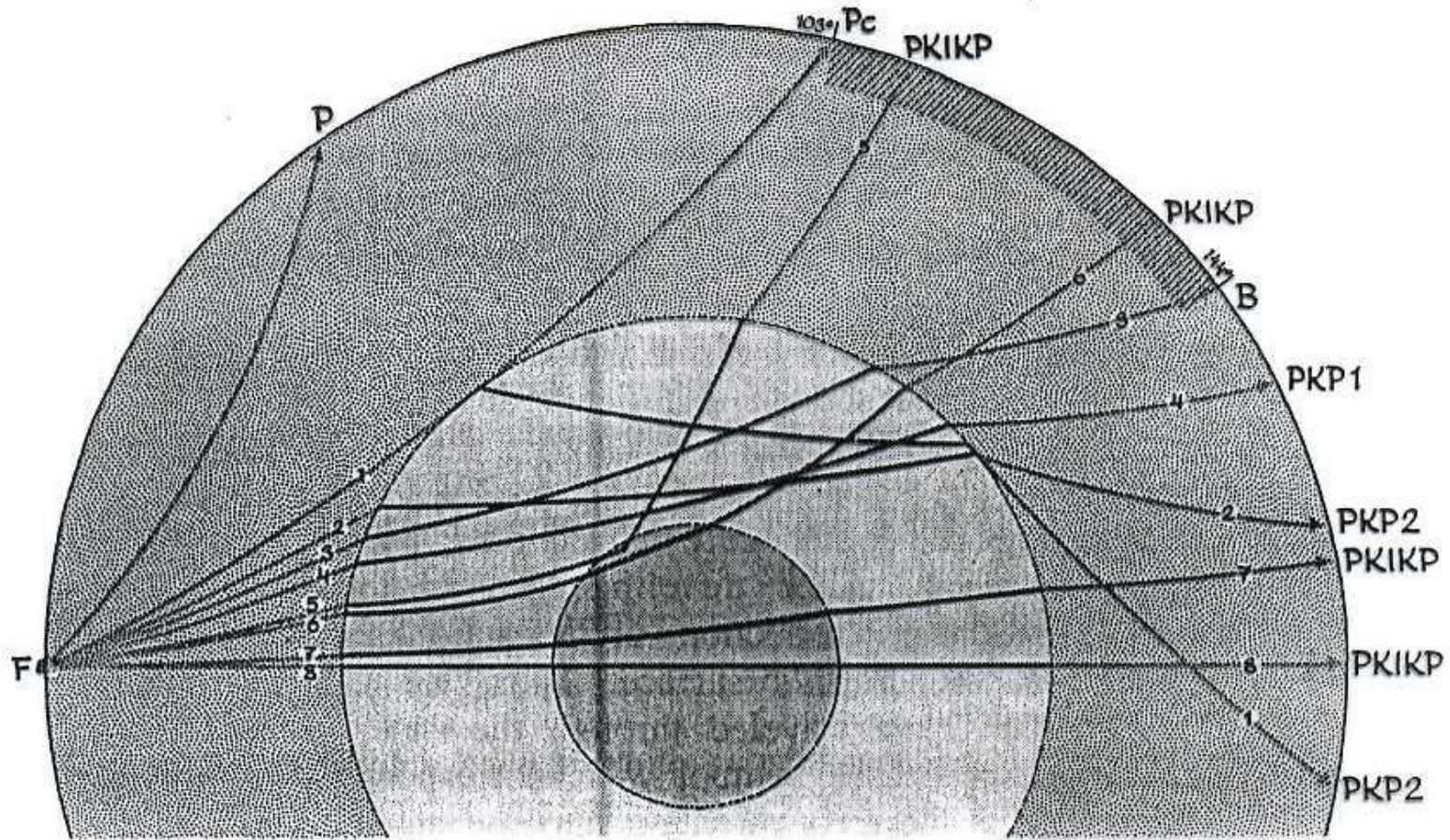


Fig. 17. Examples of propagation paths of *P* waves traveling through the Earth's core. The shadow zone between 103° and 144° is shaded and *B* denotes the caustic point. Rays are numbered in the order of increasing steepness of the initial descent. For details see the text. Conventions as for Fig. 14.

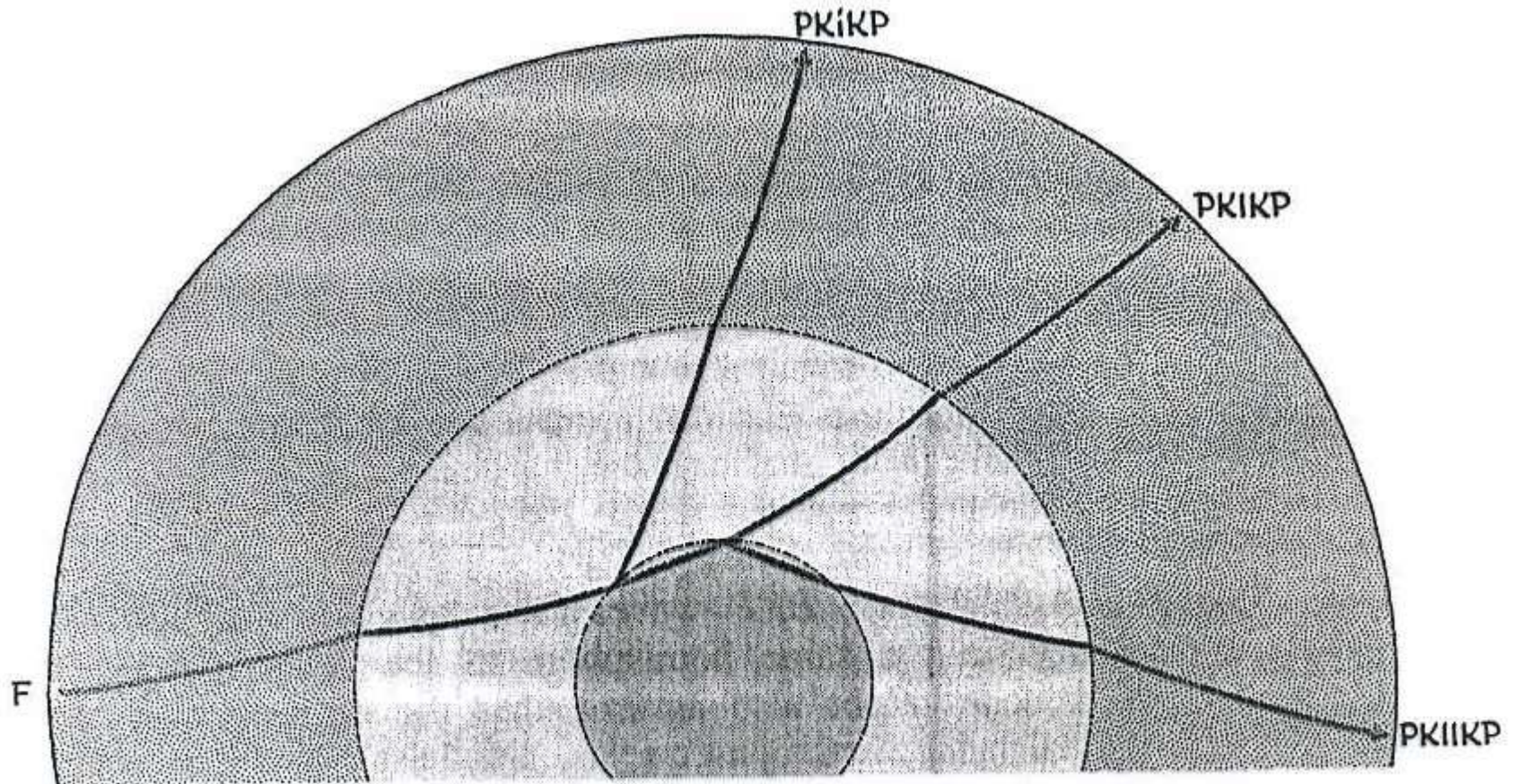


Fig. 18. Propagation paths of P waves traversing the Earth's interior from the focus F and reflecting at the outside ($PKiKP$) or inside ($PKIIKP$) of the inner core. $PKIKP$ is a P wave refracted into the outer core and through the inner core. Conventions as for Fig. 14.



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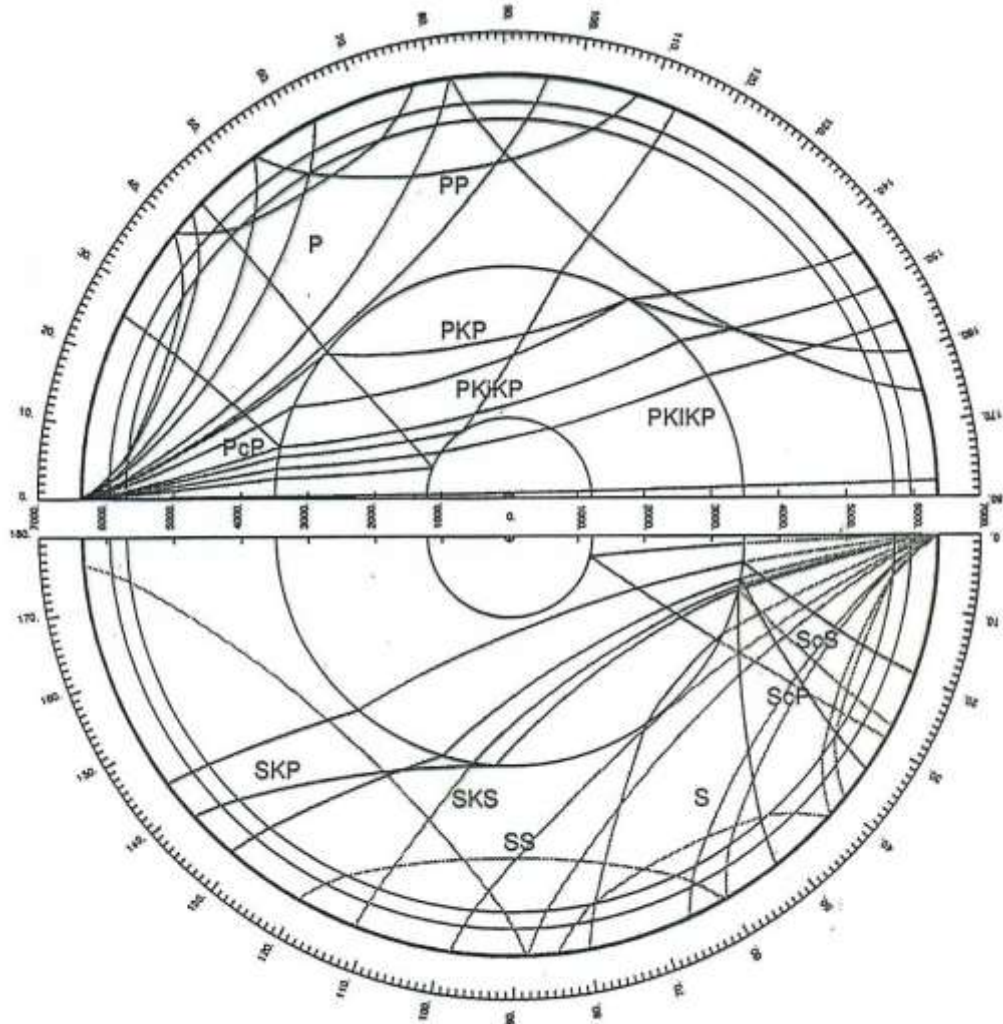
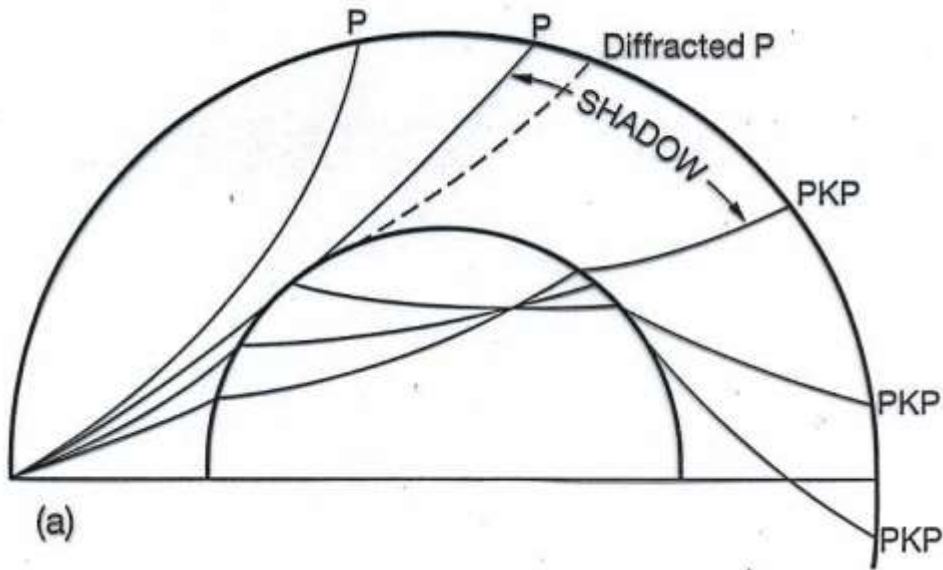


Figure 5.15. Paths of seismic rays through the Earth, illustrating their nomenclature. Figure by courtesy of B.L.N. Kennett.

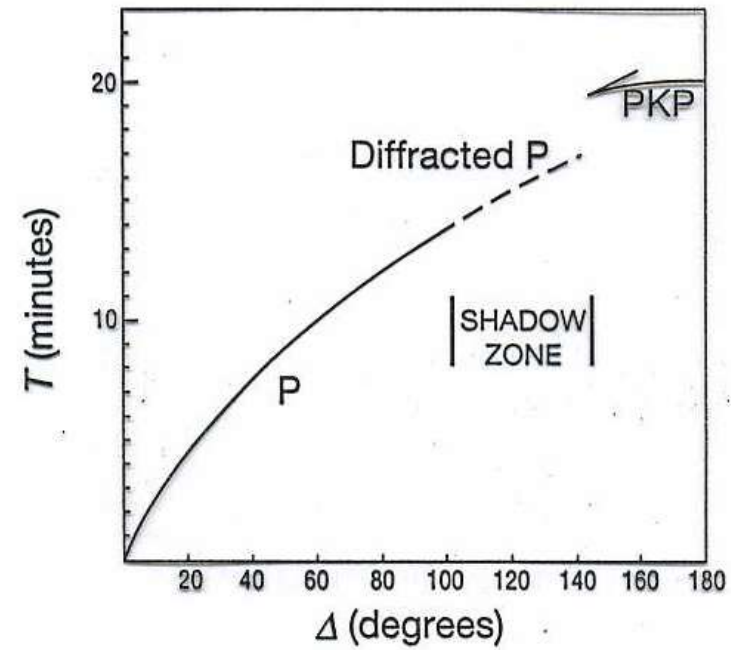


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(a)

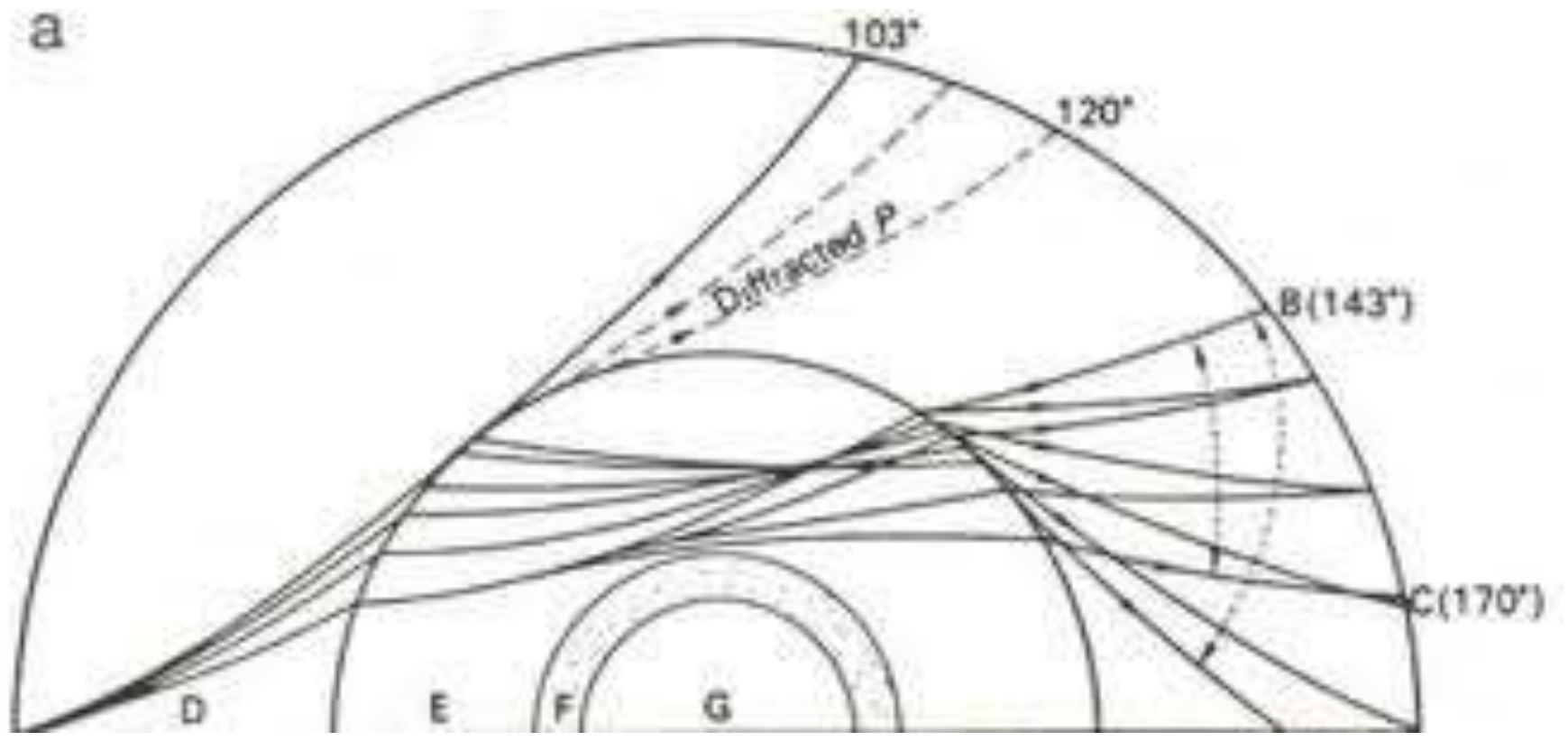


(b)



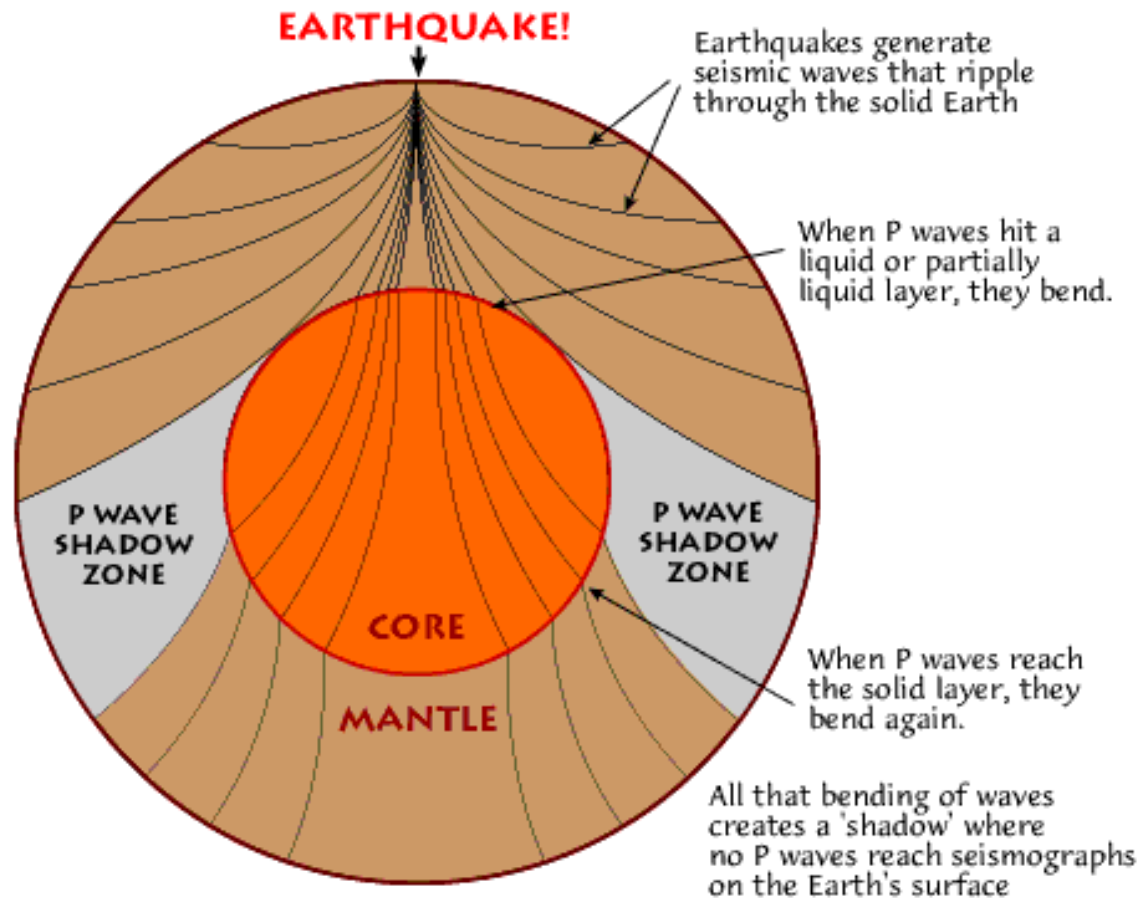
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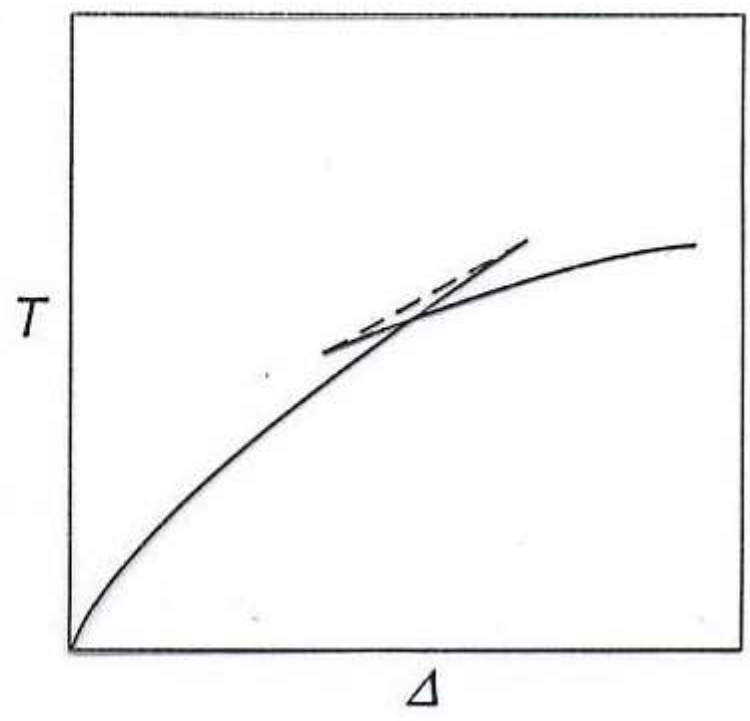
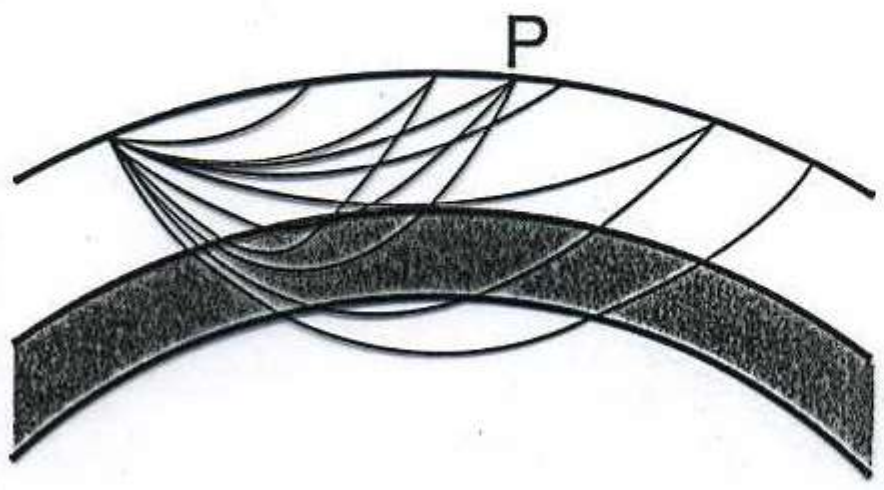
After numerous earthquakes, seismologists were able to use the pattern of P waves reaching the surface to see into the Earth. They found a layer of mostly metallic liquid we call the *outer core*.



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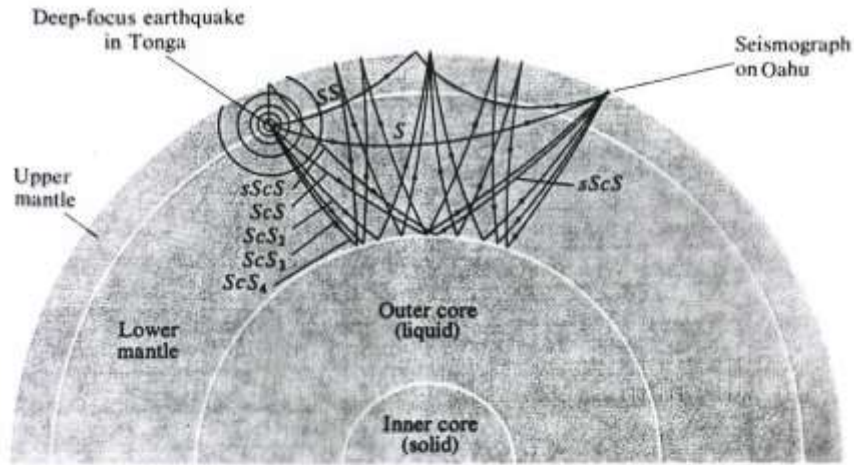
Un brusco incremento di velocità può portare alla triplicazione degli arrivi (Figura schematica!)



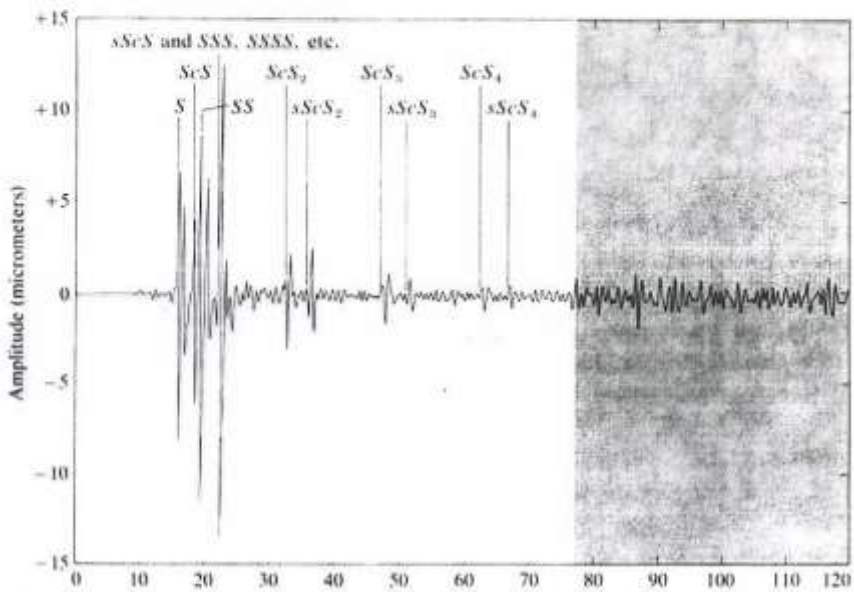


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(a)



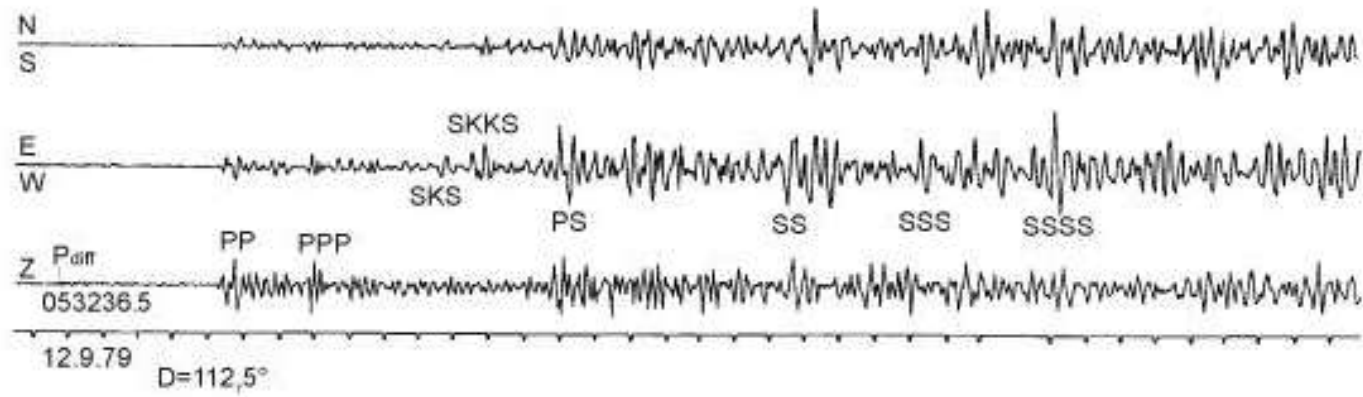
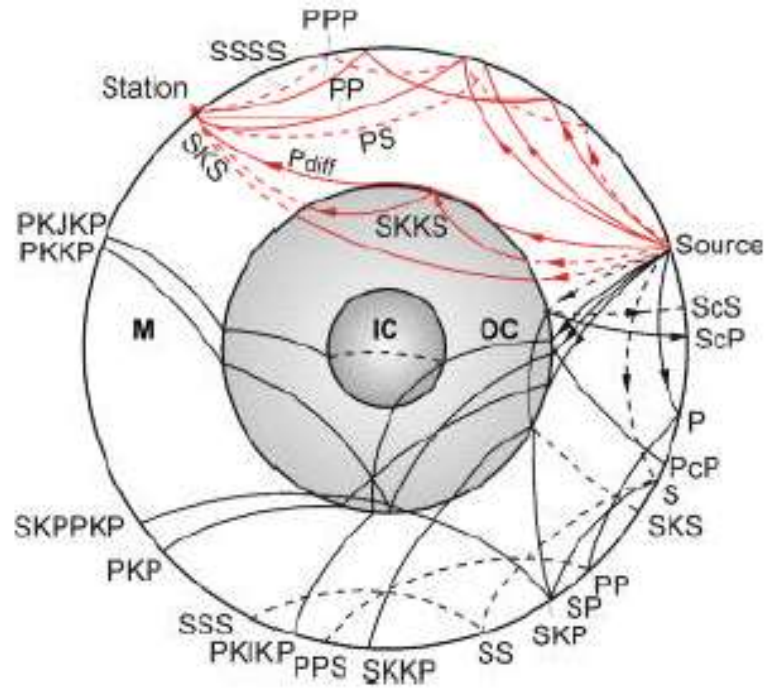
(b)

Figure 5.3 Multiple surface reflections (a) and seismogram (b) recorded at a seismographic station at Oahu, Hawaii. These waves followed a deep-focus earthquake near the South Pacific island of Tonga in October 1974. [From Thomas H. Jordan, "The Deep Structure of the Continents." Copyright © 1978 by Scientific American, Inc. All rights reserved.]



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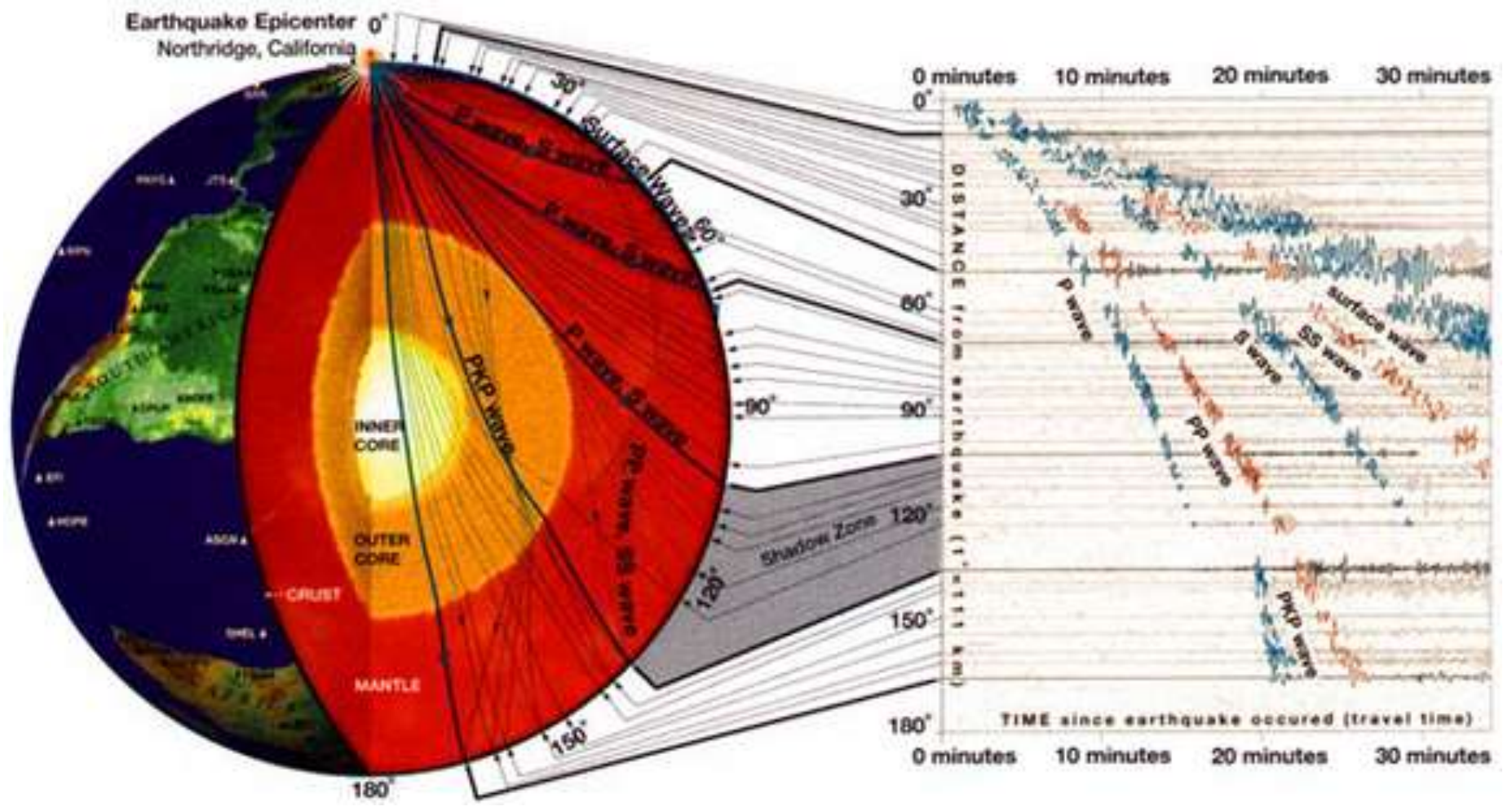
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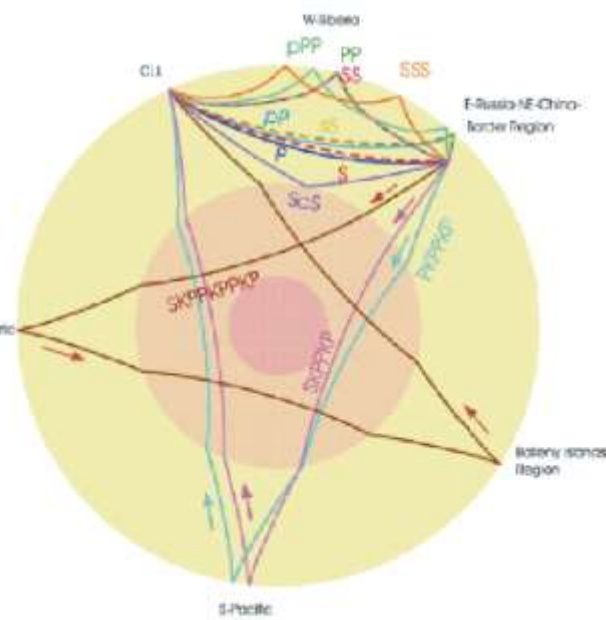
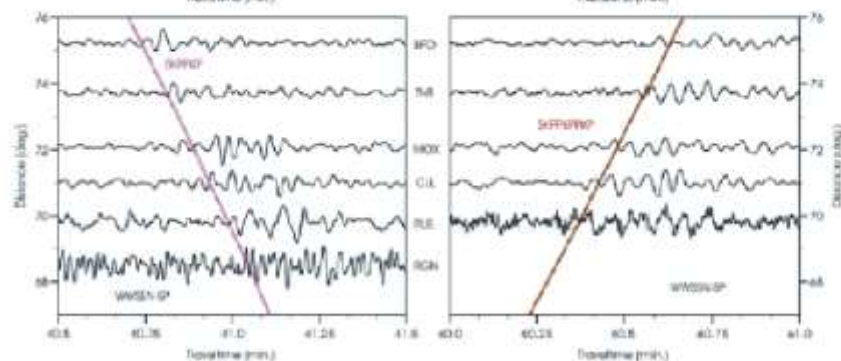
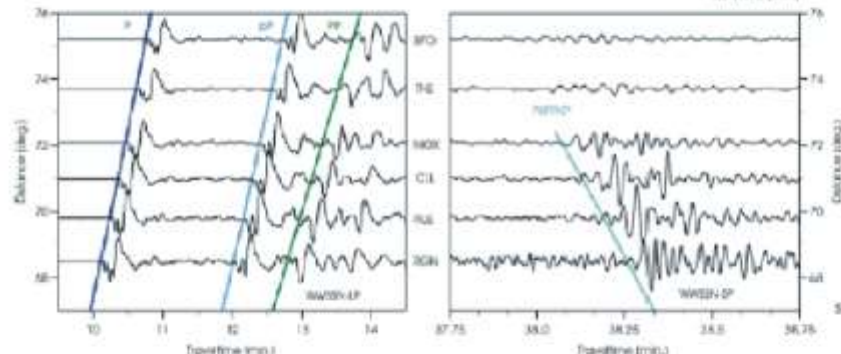
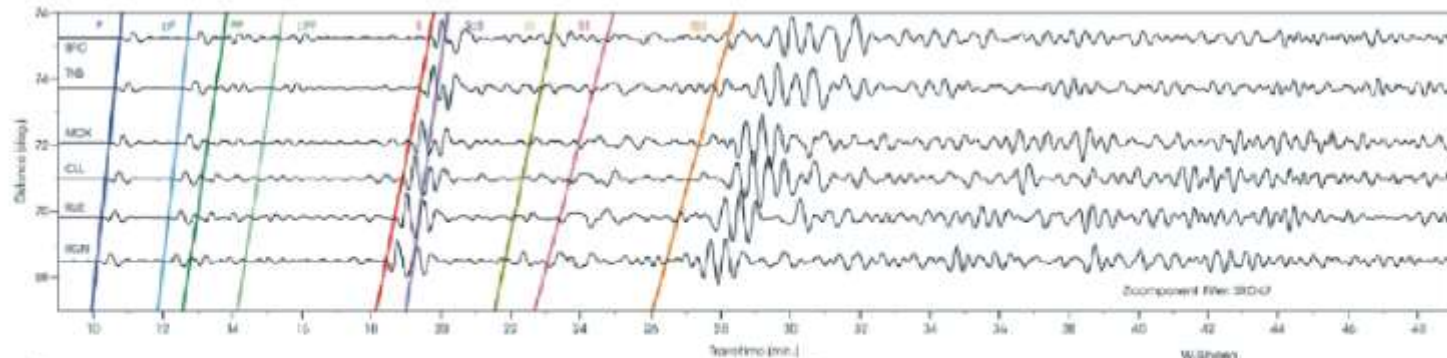
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28.06.2002 H=17:19:30.2 43.8N; 130.7E
 h=564km Mw=7.3
 E-Russia-NE-China Border Region (NEIC)



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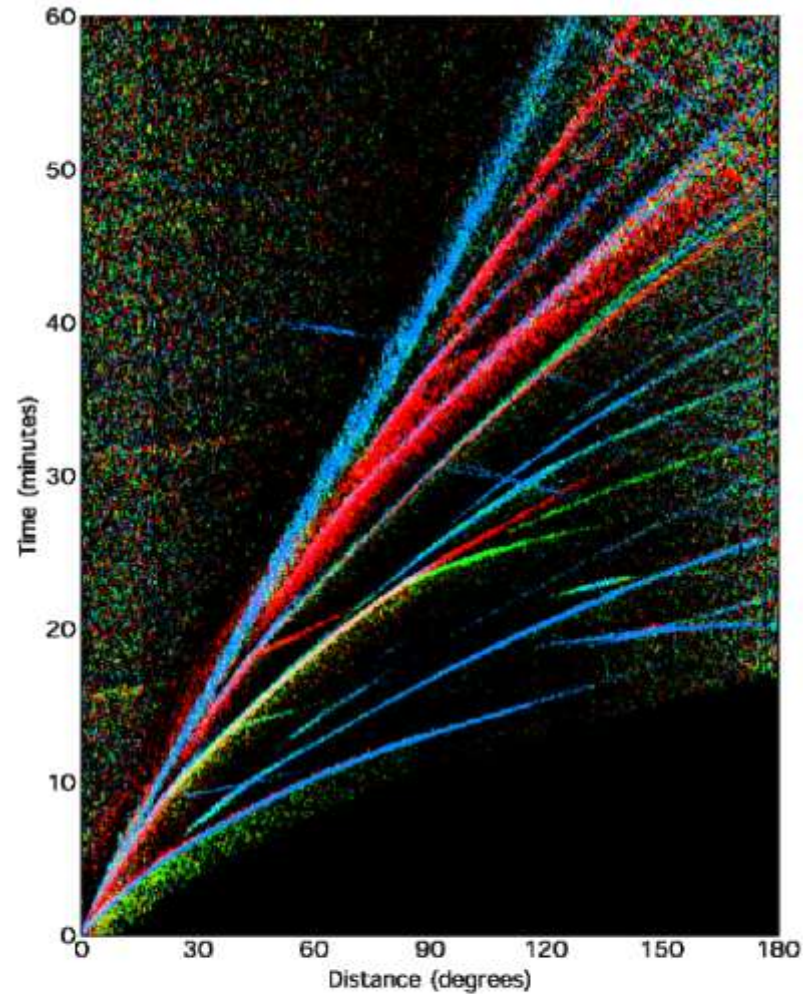


Fig. 2.49 Global travel-time curves for shallow earthquakes as produced by stacking broadband seismograms. Seismic phases are shown in different colors depending on their polarization (blue: vertical motion; green: radial-horizontal and red: transverse-horizontal) (courtesy of L. Astiz).



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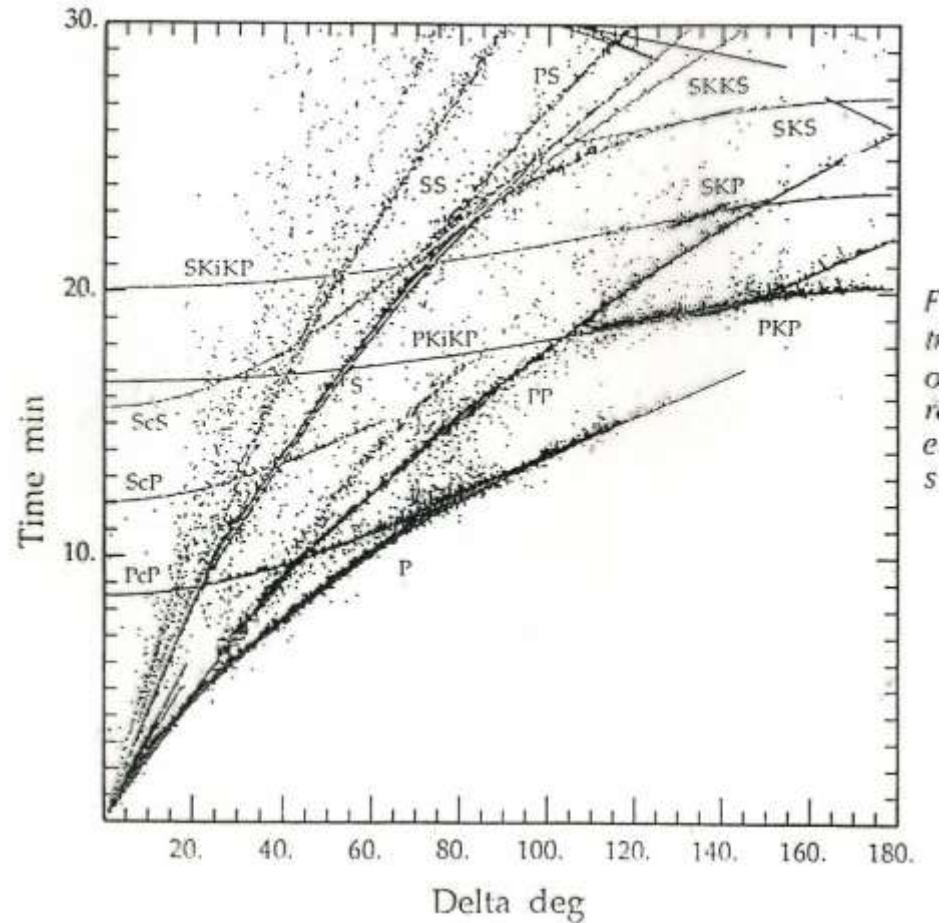
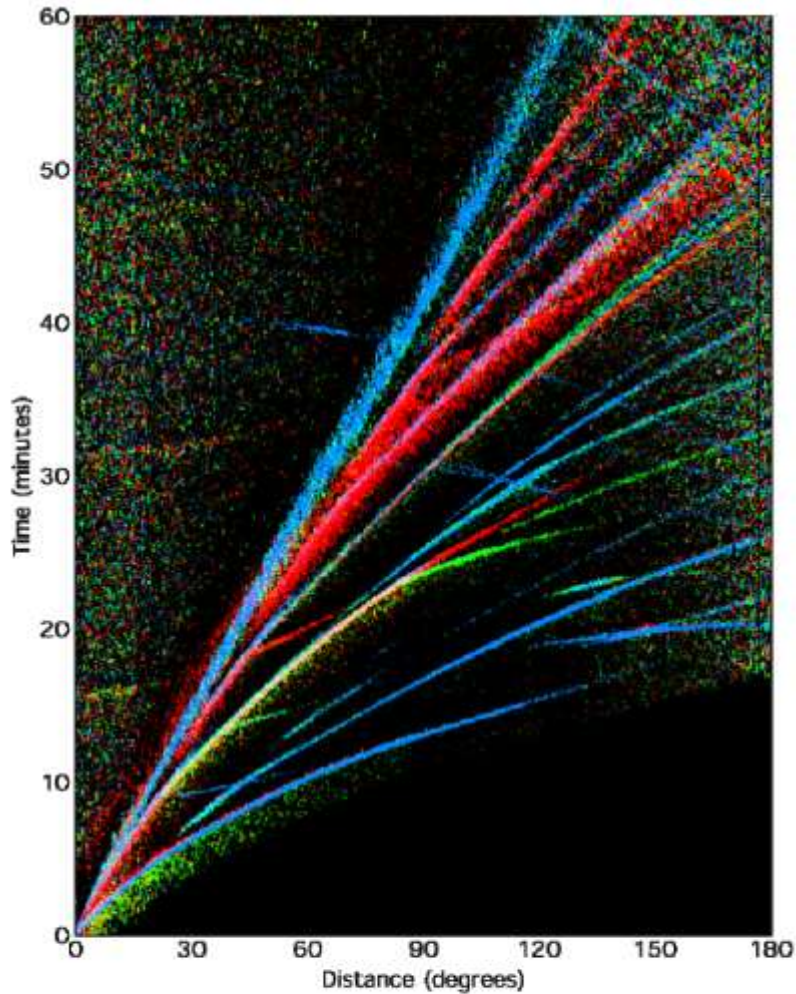
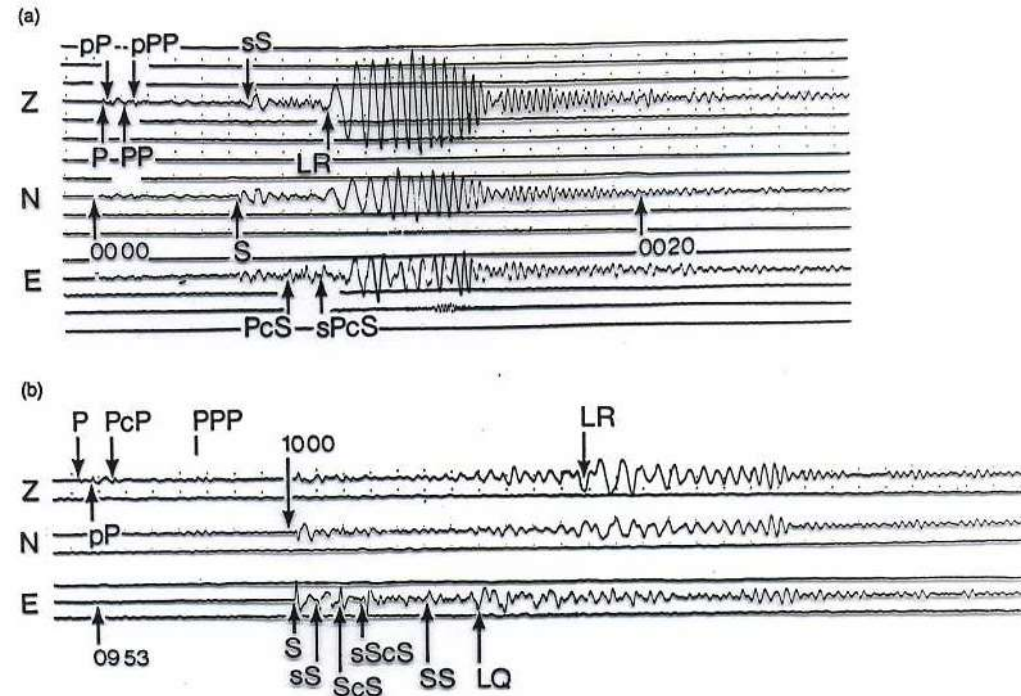
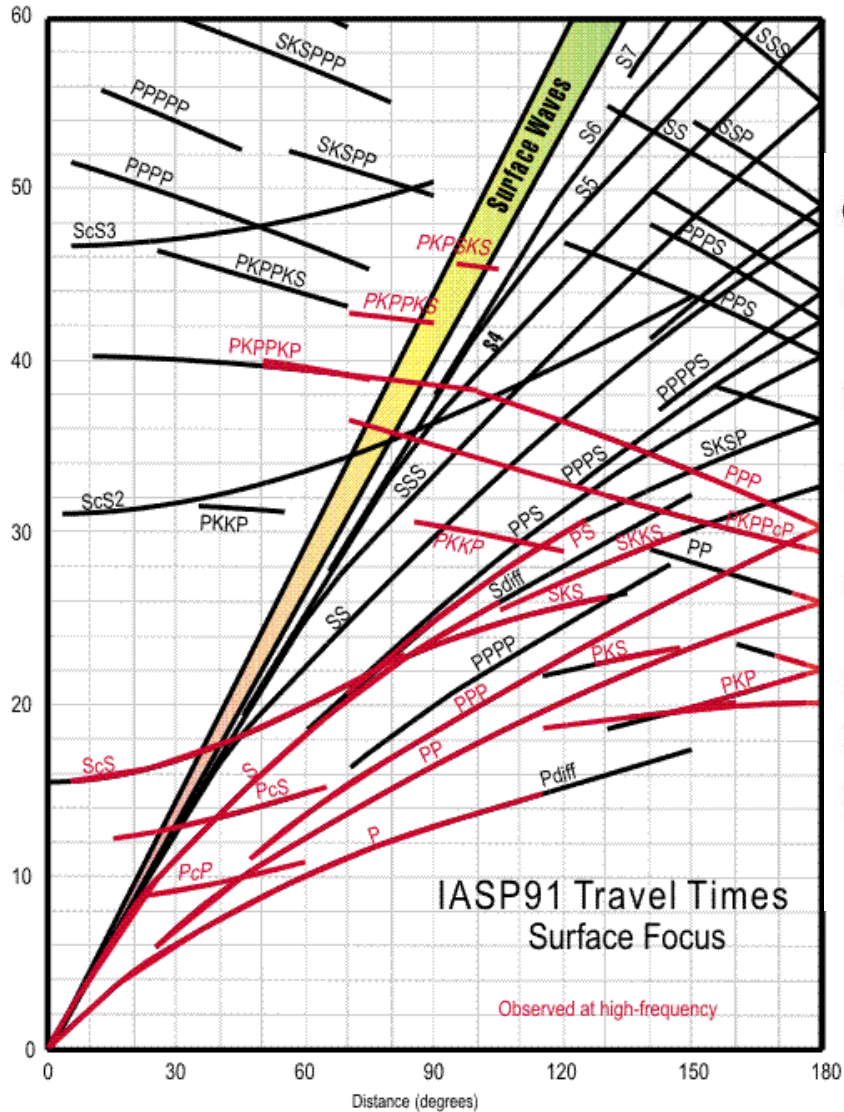


Fig. 2.49 Global travel-time curves for shallow earthquakes as produced by stacking broadband seismograms. Seismic phases are shown in different colors depending on their polarization (blue: vertical motion; green: radial-horizontal and red: transverse-horizontal) (courtesy of L. Astiz).



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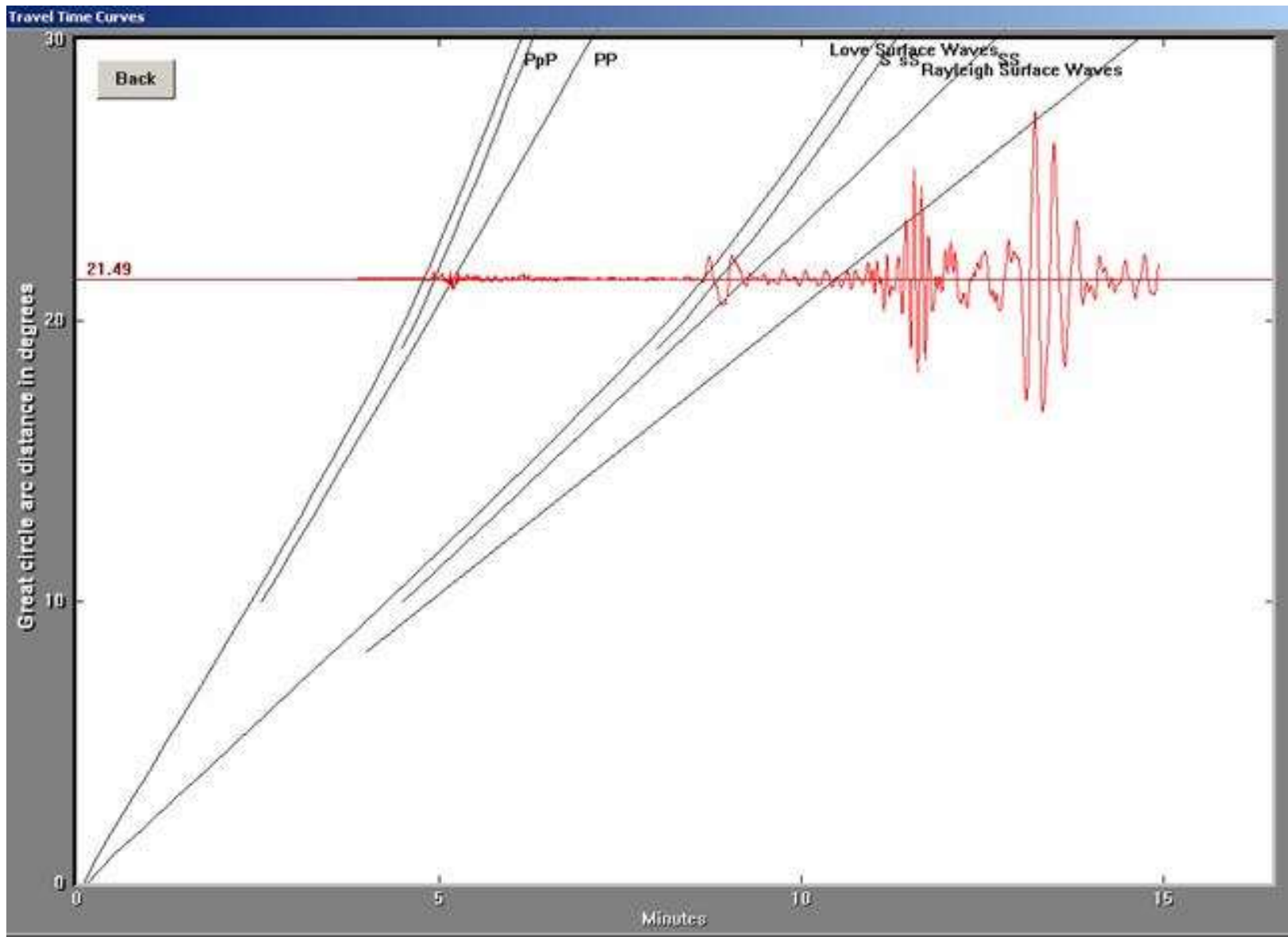
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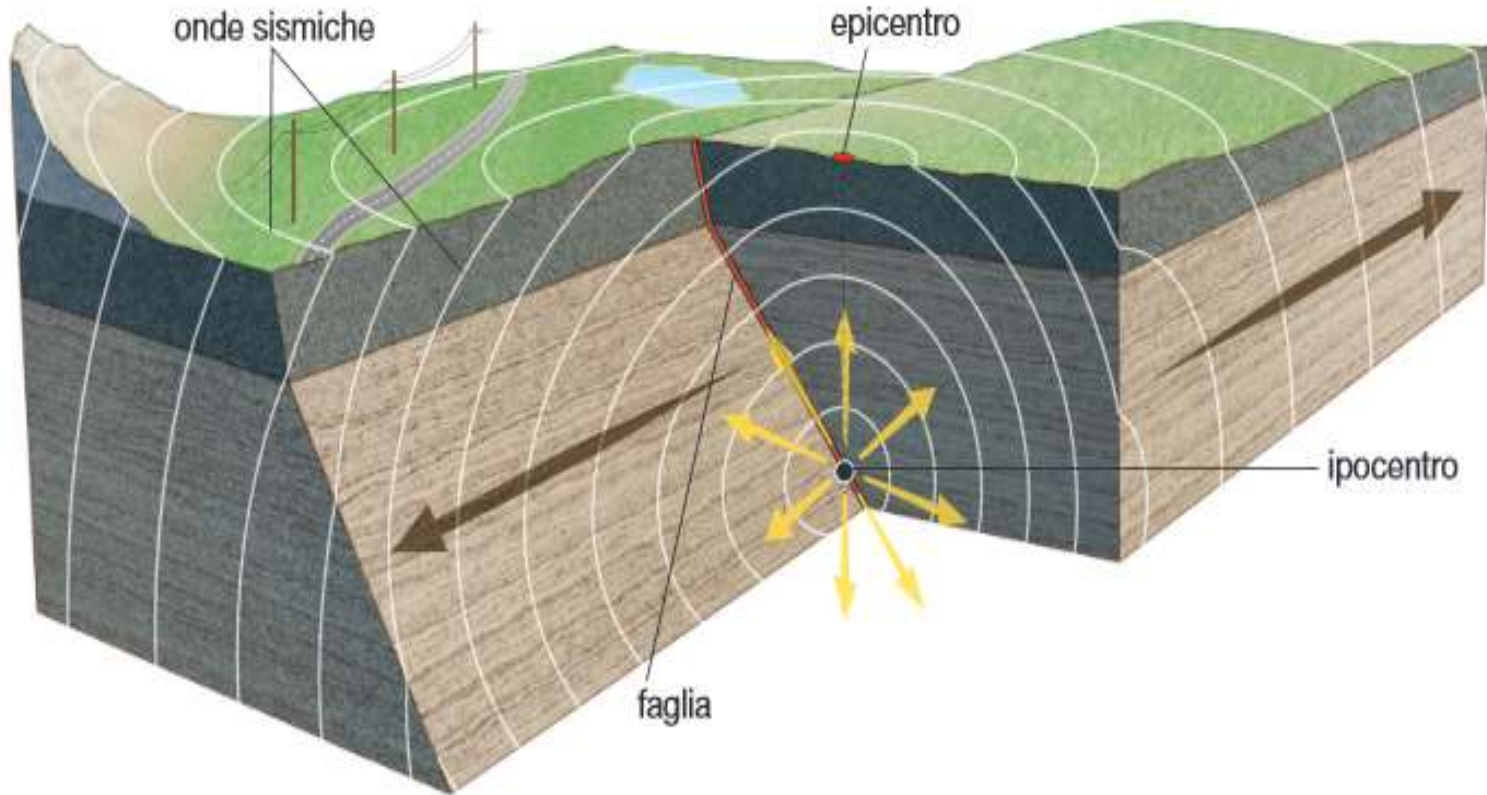
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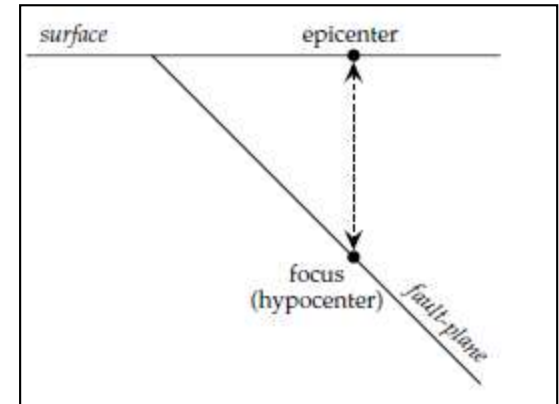
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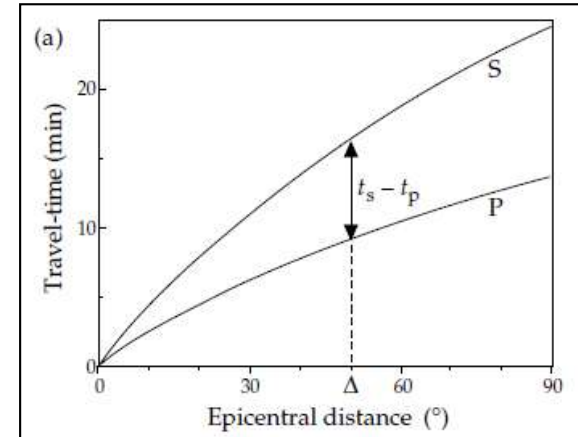
Earthquake location

- Although in fact the earthquake involves a part of the fault-plane measuring many square kilometers in area, from the point of view of an observer at a distance of hundreds or even thousands of kilometers the earthquake appears to happen at a point. This point is called the **focus** or **hypocenter** of the earthquake. It generally occurs at a focal depth many kilometers below the Earth's surface. The point on the Earth's surface vertically above the focus is called the **epicenter** of the earthquake.



- The distance of a seismic station from the epicenter of an earthquake (the epicentral distance) may be expressed in kilometers Δ_{km} along the surface, or by the angle $\Delta^\circ = (180/\pi)(\Delta_{km}/R)$ subtended at the Earth's center.

- The travel-times of P- and S-waves from an earthquake through the body of the Earth to an observer are dependent on the epicentral distance.



- The travel-time versus distance plots are not linear, because the ray paths of waves travelling to distant seismographs are curved. However, the standard seismic velocity profile of the Earth's interior is well enough known that the travel-times for each kind of wave can be tabulated or graphed as a function of epicentral distance.



Seismic Travel-time Curve

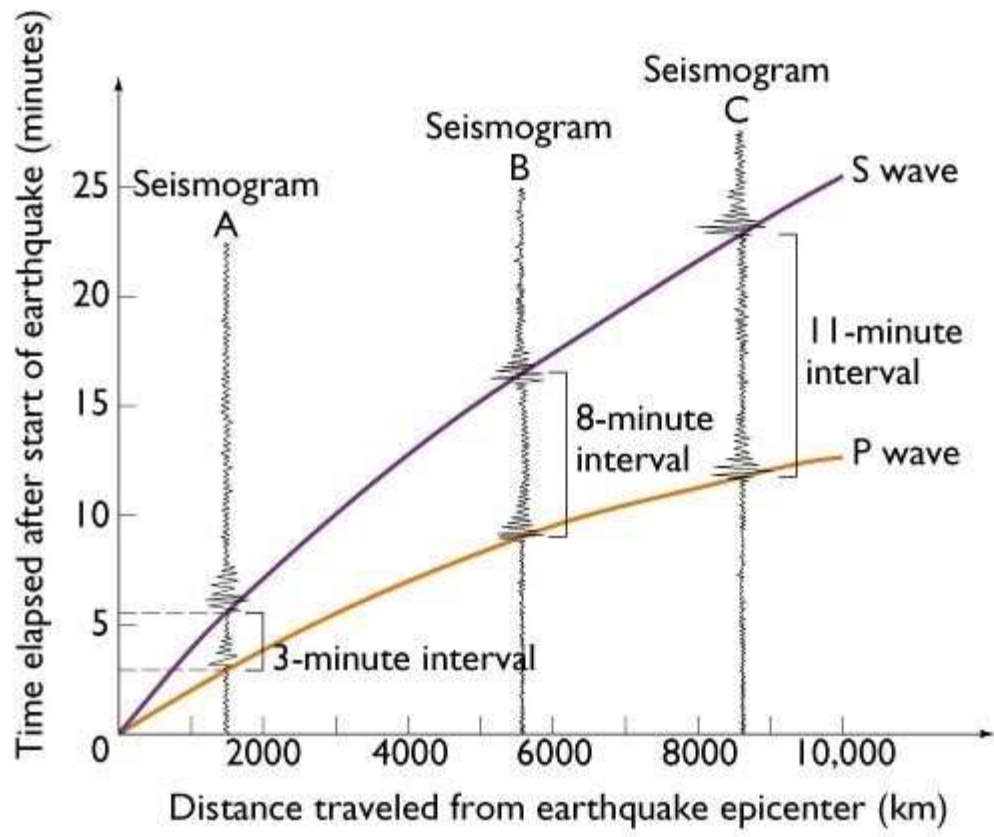


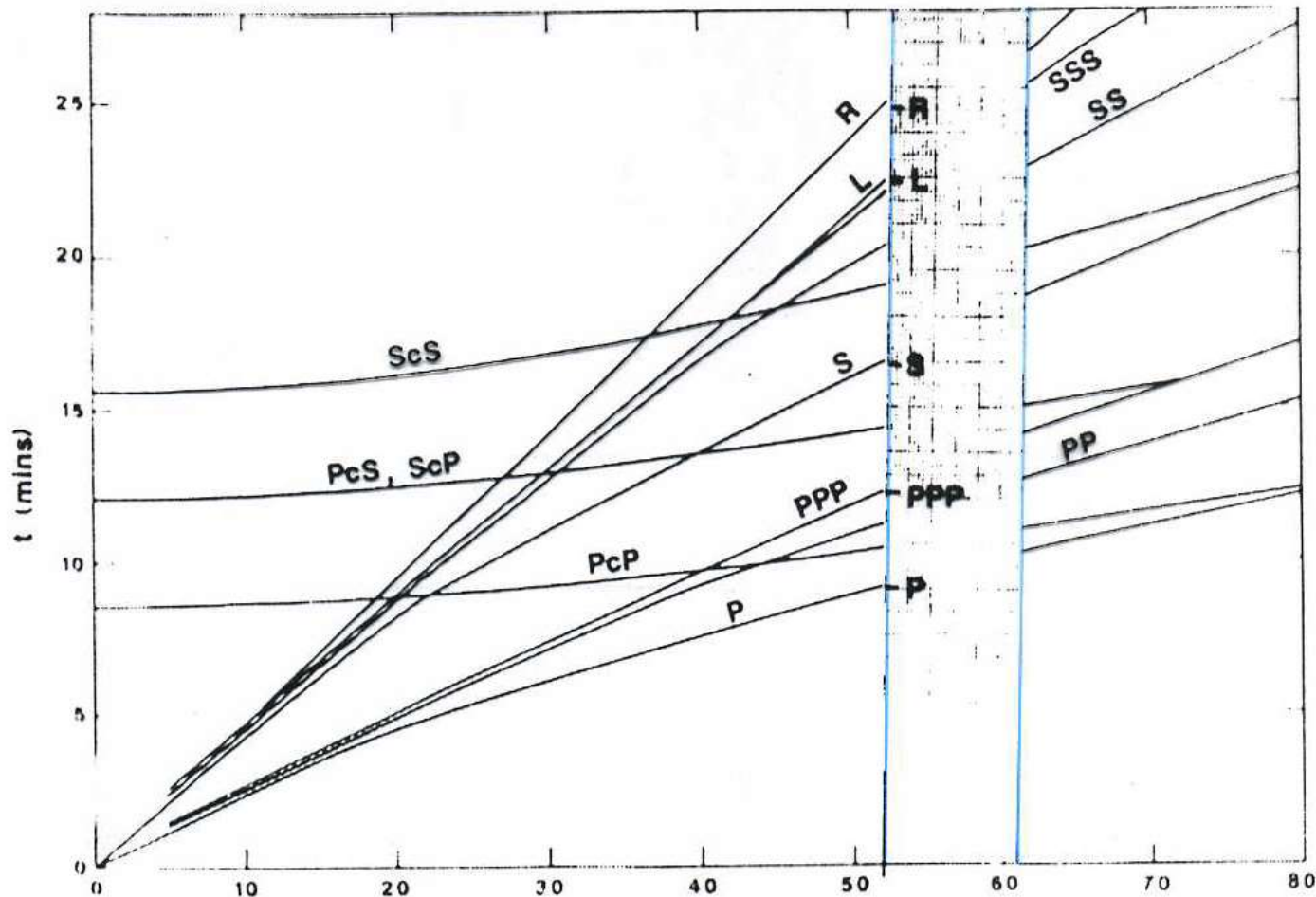
Fig. 18.9b



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Figure 4.20: Determining the distance to the earthquake having measured several arriving phases. Mark the arrival times on graph paper and lay it parallel to the time axis in Figure 4.19. Slide the graph paper along the distance axis until the marks line up.





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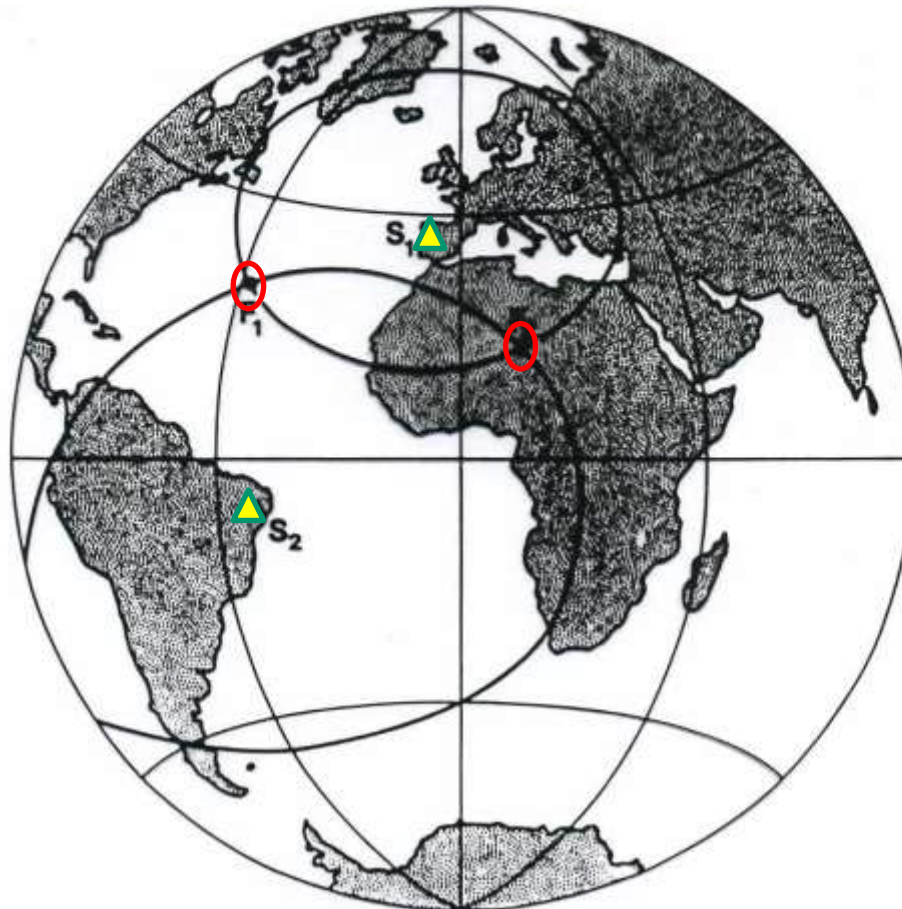
LETTURA FASI E TEMPI
D'ARRIVO

STRUMENTO	HR:MIN	DISTANZA DAL MINUTO (MM)	TEMPO (S)	TEMPO DAL 1° ARRIVO	CONVERSIONE IN MM	FASE
BUL 1	05:25	16 / 54.5	17.6	0	0	P
BUL 2	05:33	31 / 54.5	34.1	8' 16.5"	49.6	S

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Figure 4.18: S_1 and S_2 are two seismic stations. The epicentre lies on a small circle centred on a station with radius the angular distance between them. If distances can be obtained from arrivals at two stations there are two possible locations of the events, F_1 and F_2 , the intersection points of the two circles. A third station can be used to determine which of the two points is the true epicentre.





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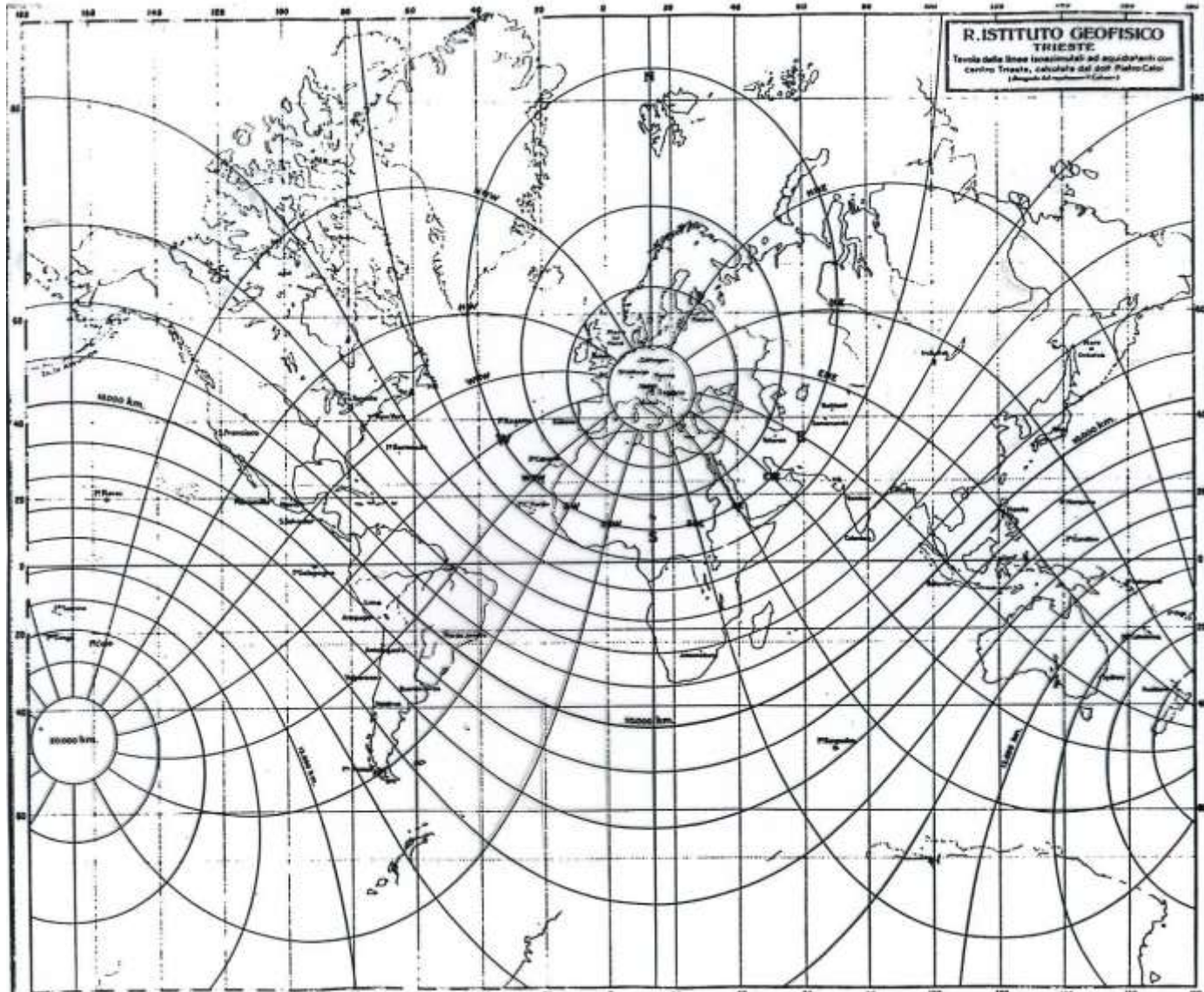
Fig. 1. - Before the widespread use of computers, earthquakes were located manually on globes such as the one pictured here. Epicenters were determined by triangulation, drawing arcs with a large compass centered on the stations that recorded the earthquake. The locations obtained using these crude methods are still the only available estimates for most of the earthquakes located before the 1960s. Photo courtesy of the National Earthquake Information Center archives.





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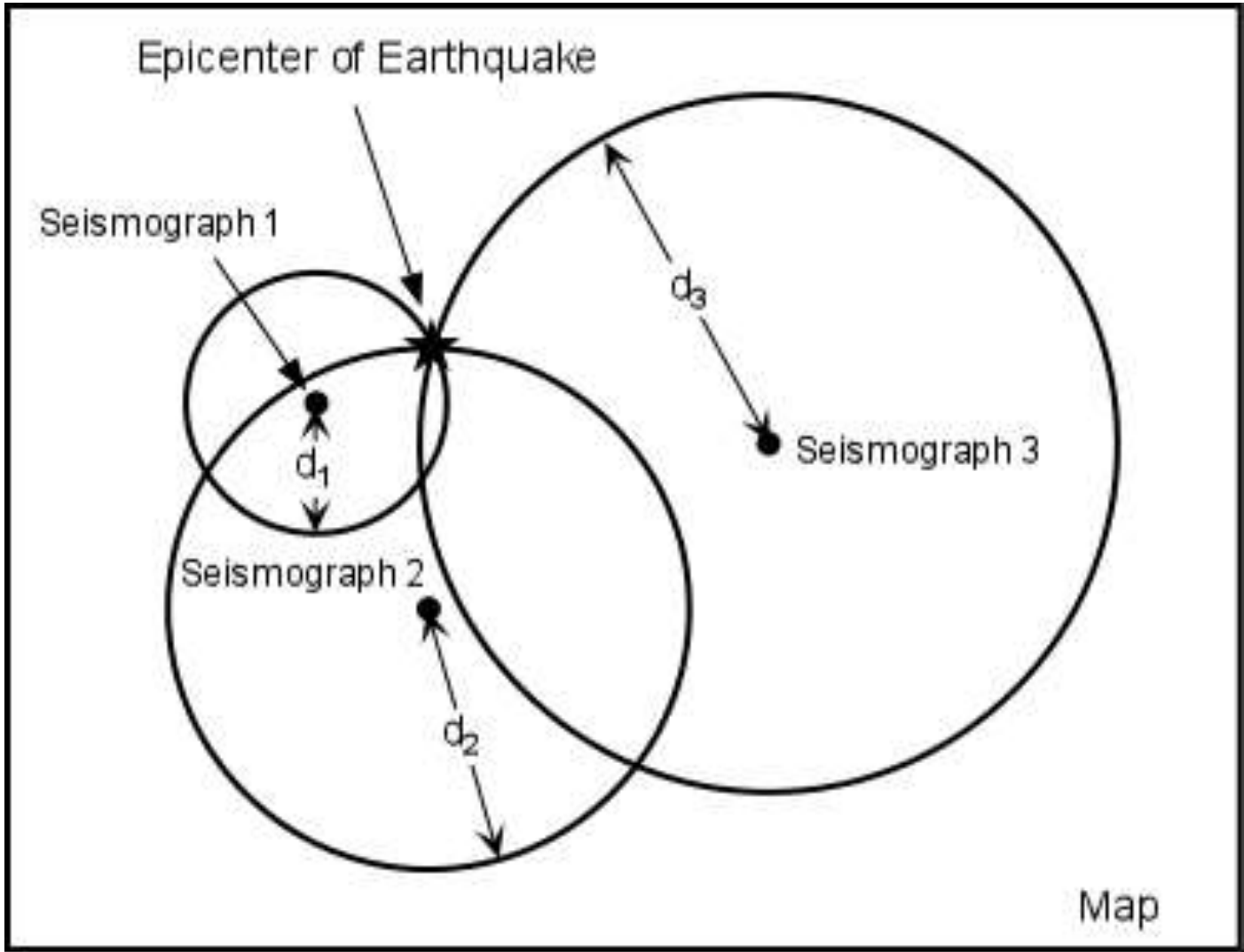
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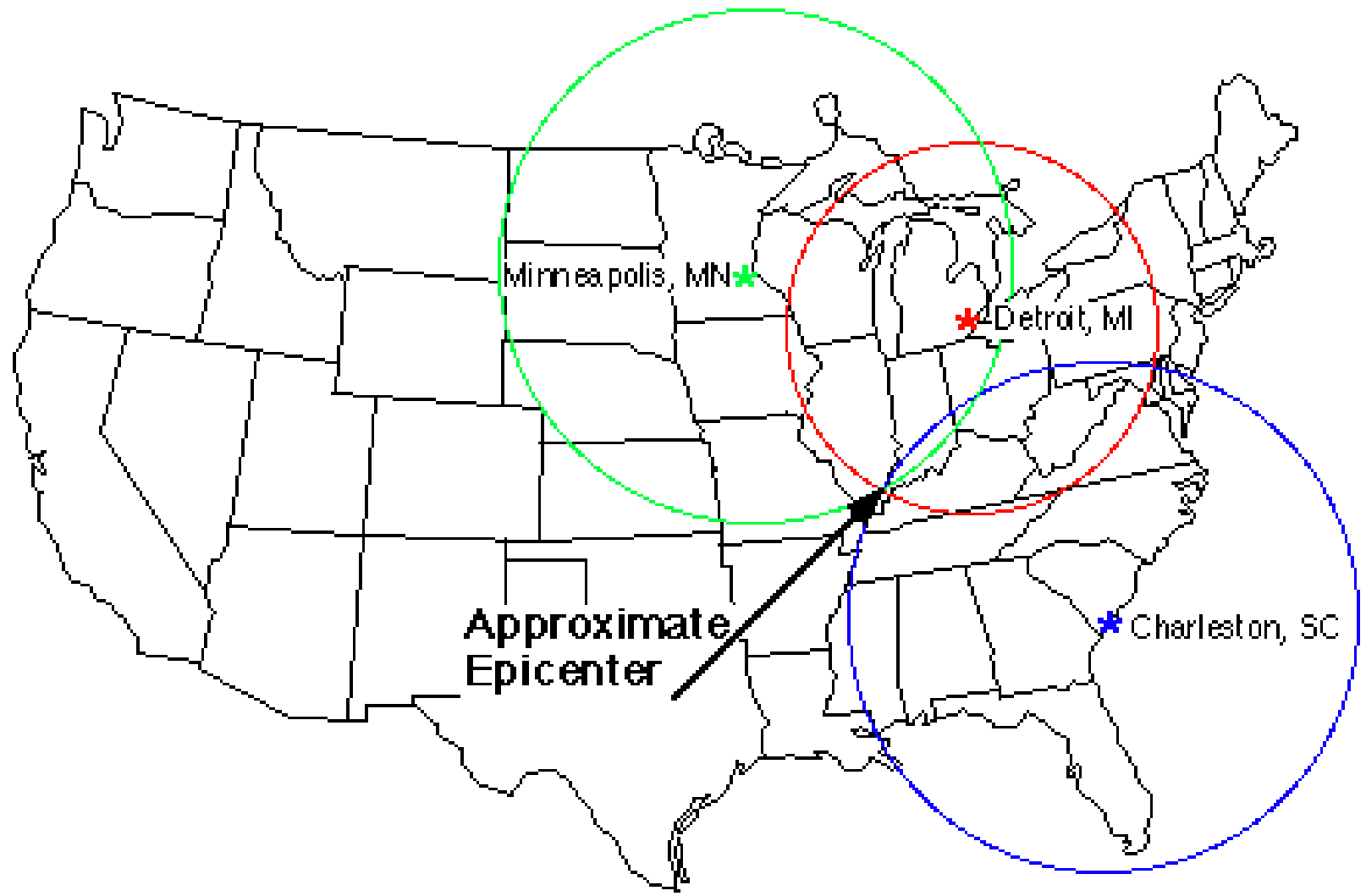
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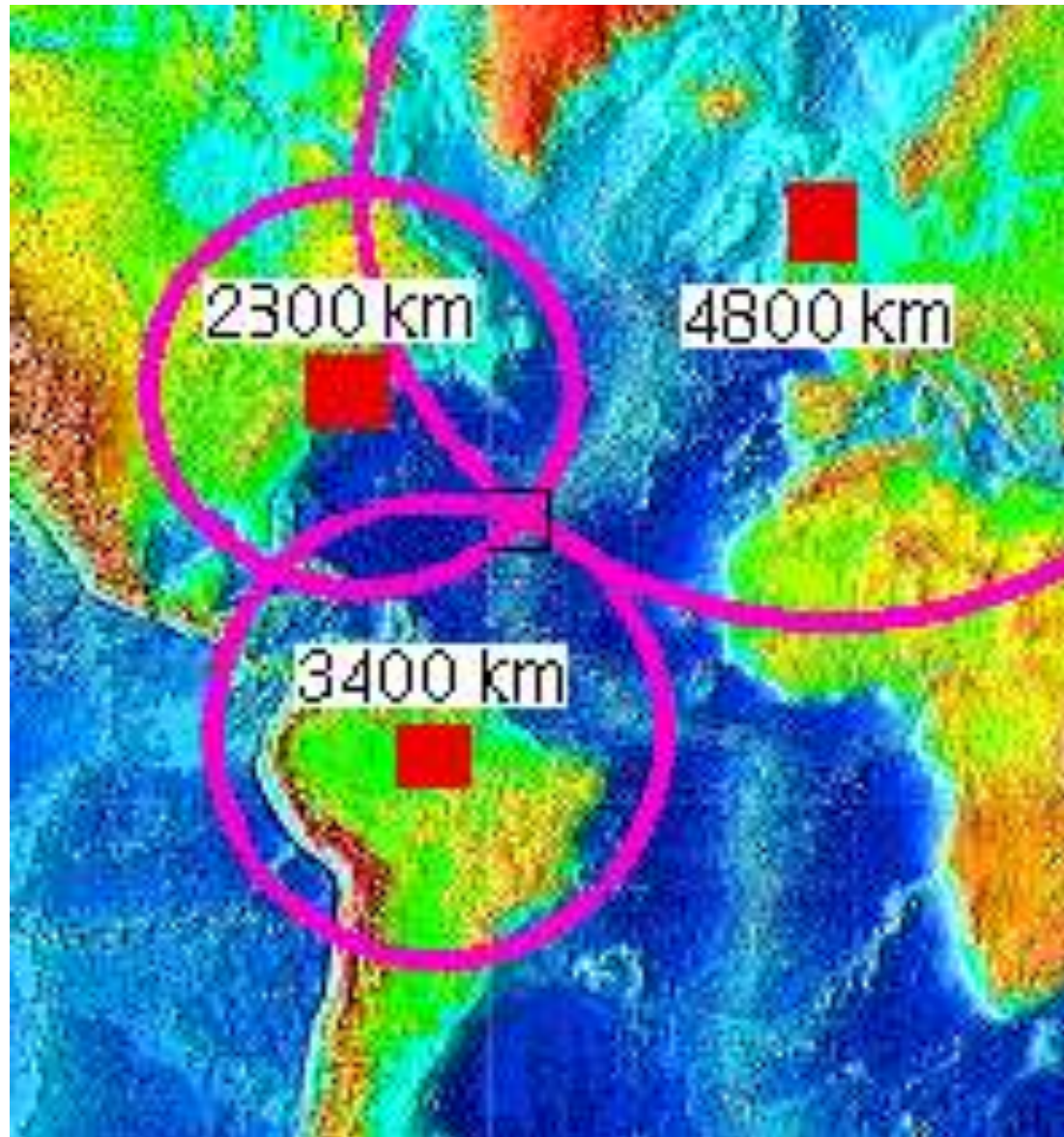
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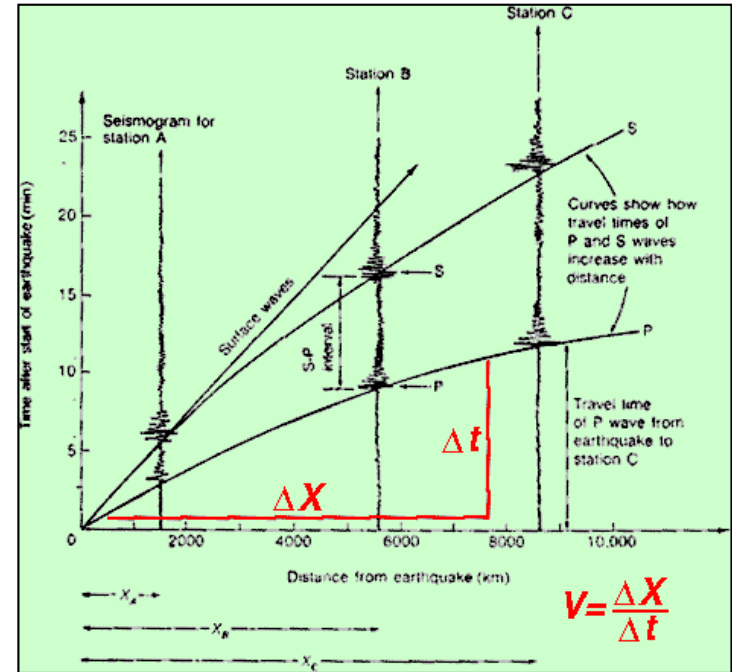
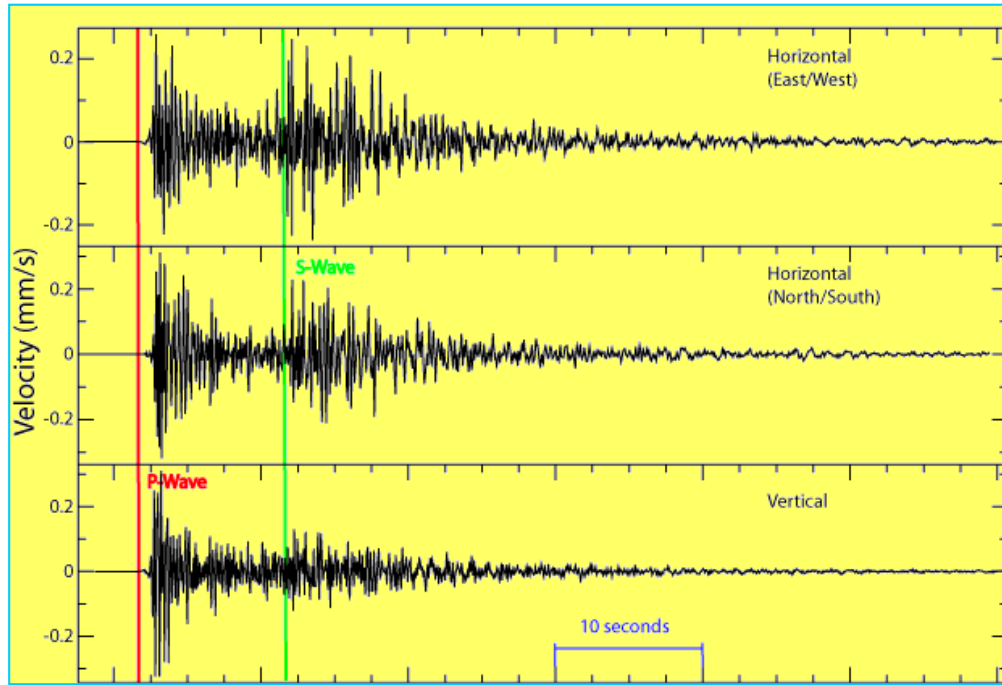
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- In computing epicentral distance from earthquake data the total travel-time is not at first known, because an observer is rarely at the epicenter to record the exact time of occurrence t_0 of the earthquake. However, the difference in travel-times for P- and S-waves ($t_s - t_p$) can be obtained directly from the seismogram; it increases with increasing epicentral distance.



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PARAMETRICI CINEMATICI DELLA SORGENTE E LOCALIZZAZIONE DEI TERREMOTI

Diremo **IPOCENTRO** o anche **FUOCO DEL TERREMOTO** il punto nella Terra che per primo emana onde sismiche nel corso di un evento sismico (nucelazione del terremoto). Esso è determinato da tre parametri: **longitudine, latitudine e profondità**.

Dicesi **EPICENTRO** del terremoto il punto sulla superficie terrestre che si trova sulla verticale sopra l'ipocentro.

Come vedremo, comunque, i terremoti non avvengono in un punto, ma lungo i piani di faglia. Nel caso di grossi terremoti, le dimensioni fisiche della sorgente (faglia) possono essere di diverse centinaia di km e quindi l'ipocentro può in linea di principio essere localizzato in un punto qualunque della superficie di rottura. Poiché la localizzazione di un terremoto è determinata attraverso i tempi di arrivo delle fasi sismiche generate dalla fase iniziale del processo di rottura, la localizzazione in generale coinciderà (a meno di errori di analisi) con la zona della sorgente dove il terremoto ha avuto origine. In particolare questo è vero per le fasi P ed S poiché la velocità con cui la rottura si propaga lungo la superficie della faglia è minore delle velocità di queste due fasi.



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Dicesi **TEMPO ORIGINE** il tempo t_0 in cui avviene l'evento. Per via dei diversi fusi orari nel mondo, il tempo origine va espresso in termini di tempo di **Greenwich medio (tempo assoluto, UTC, GMT)**.

Per localizzare una sorgente sismica, bisogna pertanto determinare **4** quantità. Le informazioni che abbiamo a disposizione sono rappresentate dai tempi di arrivo delle onde (**P ed S**) alle varie stazioni.

Esistono vari metodi per localizzare i terremoti, a seconda se conosciamo o no le velocità della struttura della zona o meno, se abbiamo a disposizione i tempi di arrivo delle onde P od anche delle onde S, se abbiamo una stazione a tre componenti o più stazioni ad una componente (verticale).

Ovviamente, il numero minimo di dati necessari è **4**.



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METODO DI WADATI

Assunzioni:

Sorgente puntiforme

Mezzo omogeneo (velocità costanti)

Stazioni vicine (distanze dell'ordine di grandezza della profondità del terremoto)

La propagazione delle onde è data da:

$$R = V(t - t_0)$$

Dove R è la distanza ipocentrale, V è la velocità dell'onda, t è il tempo di arrivo dell'onda alla stazione, t_0 è il tempo origine del terremoto.

$$R = V_P (t_P - t_0)$$

$$R = V_S (t_S - t_0)$$

$$V_P t_P - V_P t_0 = V_S t_S - V_S t_0$$



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$$V_P t_P - V_P t_0 = V_S t_S - V_S t_0$$

$$t_0 = \frac{V_P t_P - V_S t_S}{V_P - V_S}$$

Sostituendo

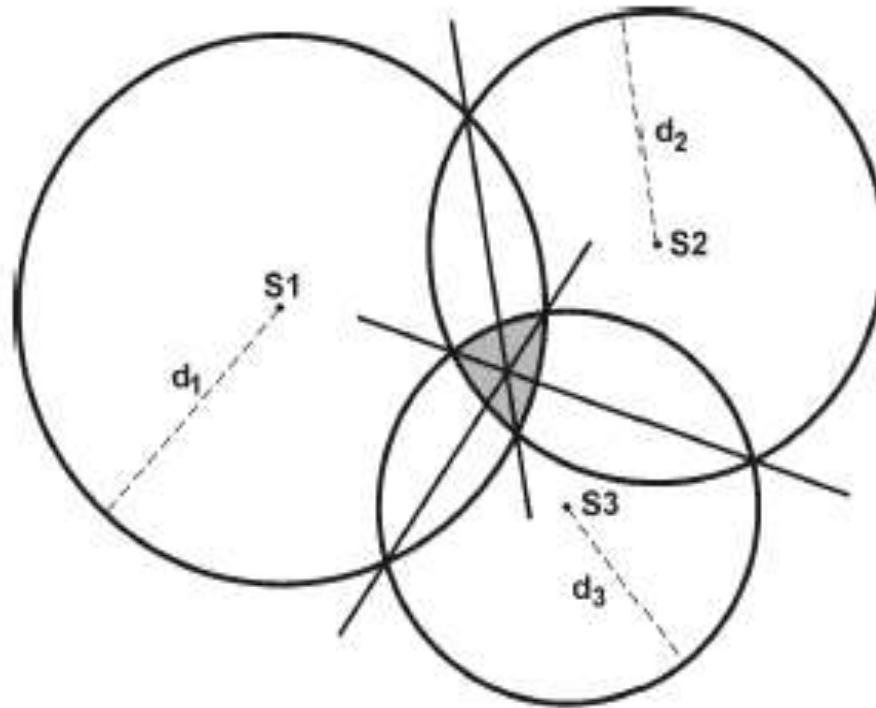
$$R = V_P \left(t_P - \frac{V_P t_P - V_S t_S}{V_P - V_S} \right)$$

$$R = \frac{V_P V_S}{V_P - V_S} (t_S - t_P)$$



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$$d_i = \frac{V_P V_S}{V_P - V_S} (t_S - t_P)$$

Per determinare l'epicentro di un evento nel caso di velocità V_P e V_S siano conosciute nell'area e si abbiano a disposizione i tempi di arrivo delle onde P ed S a tre stazioni, si può calcolare la distanza d_i dell'evento da ogni stazione e determinare l'epicentro.



DETERMINAZIONE DEL TEMPO ORIGINE

Eliminando dalle equazioni: $R = V_P (t_P - t_0)$ e $R = V_S (t_S - t_0)$ la quantità R ottenuto

$$\frac{t_S - t_0}{t_P - t_0} = \frac{V_P}{V_S}$$

Assumiamo che all'interno della rete di stazioni il rapporto V_P/V_S sia costante e ponendo $T_S = t_S - t_0$ e $T_P = t_P - t_0$, otteniamo moltiplicando per T_P e sottraendo T_P

$$t_S - t_P = \left(\frac{V_P}{V_S} - 1 \right) t_P - \left(\frac{V_P}{V_S} - 1 \right) t_0$$

Quindi, il tempo $T_{PS} = t_S - t_P$ è una funzione lineare di t_P . Le incognite sono il rapporto V_P/V_S ed il tempo origine t_0 . Per determinarli abbiamo bisogno dati di almeno due stazioni.



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Riportando in grafico (in ascisse i tempi t_p ed in ordinata i tempi $t_s - t_p$ delle corrispondenti stazioni) i punti e trovando la retta che meglio approssima il loro andamento, si ottiene il tempo t_0 come intercetta della retta con l'asse delle ascisse.

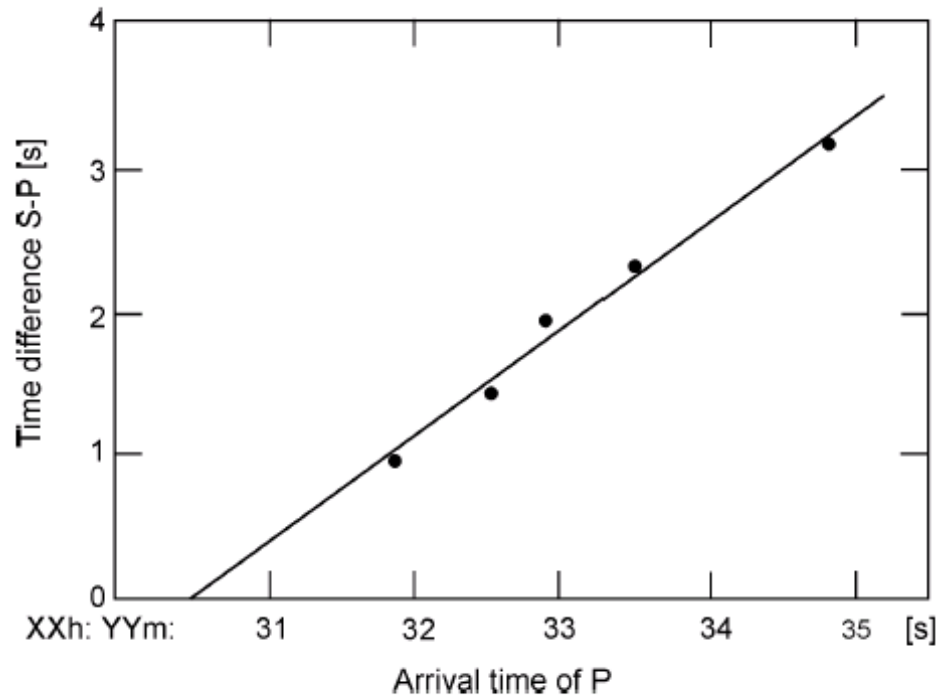


Figure 3 An arbitrary example of a Wadati diagram. The intercept of the best fitting line through the data with the x-axis gives the origin time OT. In the given case, the slope of the line is 0.72 so the v_p/v_s ratio is 1.72. This misfit of the data with a straight line indicates model and/or data reading errors.



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Il rapporto $\frac{V_P V_S}{V_P - V_S}$ è costante mentre R , t_p e t_s variano da stazione a stazione.

Il rapporto tra le distanze ipocentrali per le stazioni i e k sarà:

$$\frac{R_i}{R_k} = \frac{t_S - t_P}{t_P - t_P} = m_{ik}$$

Il significato geometrico del rapporto è il luogo dei punto per cui le distanze R_i e R_j da due punti fissi hanno rapporto costante m_{ij} , è la superficie di una sfera con centro sulla retta congiungente i due punti.

Ad ogni coppia di stazioni corrisponde una sfera. L'intersezione di due sfere è un cerchio. L'intersezione del cerchio con una terza sfera determina due punti situati simmetricamente rispetto al piano passante per i centri dei tre cerchi. Il punto al di sotto del piano (i.e. Nella terra) è l'ipocentro.

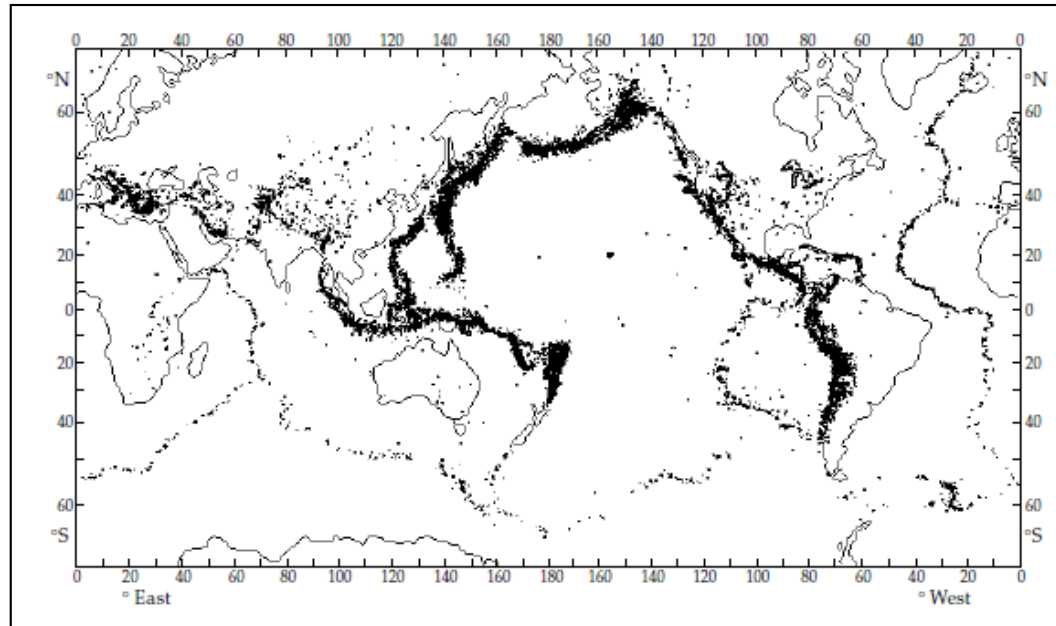
Per determinarlo abbiamo bisogno pertanto di **almeno quattro stazioni**.



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- The main reason for the intersection triangle or polygon is, however, that the seismic rays travel to the seismograph from the **focus**, and not from the epicenter. The focal depth of the earthquake, **d**, which may be up to several hundred kilometers.

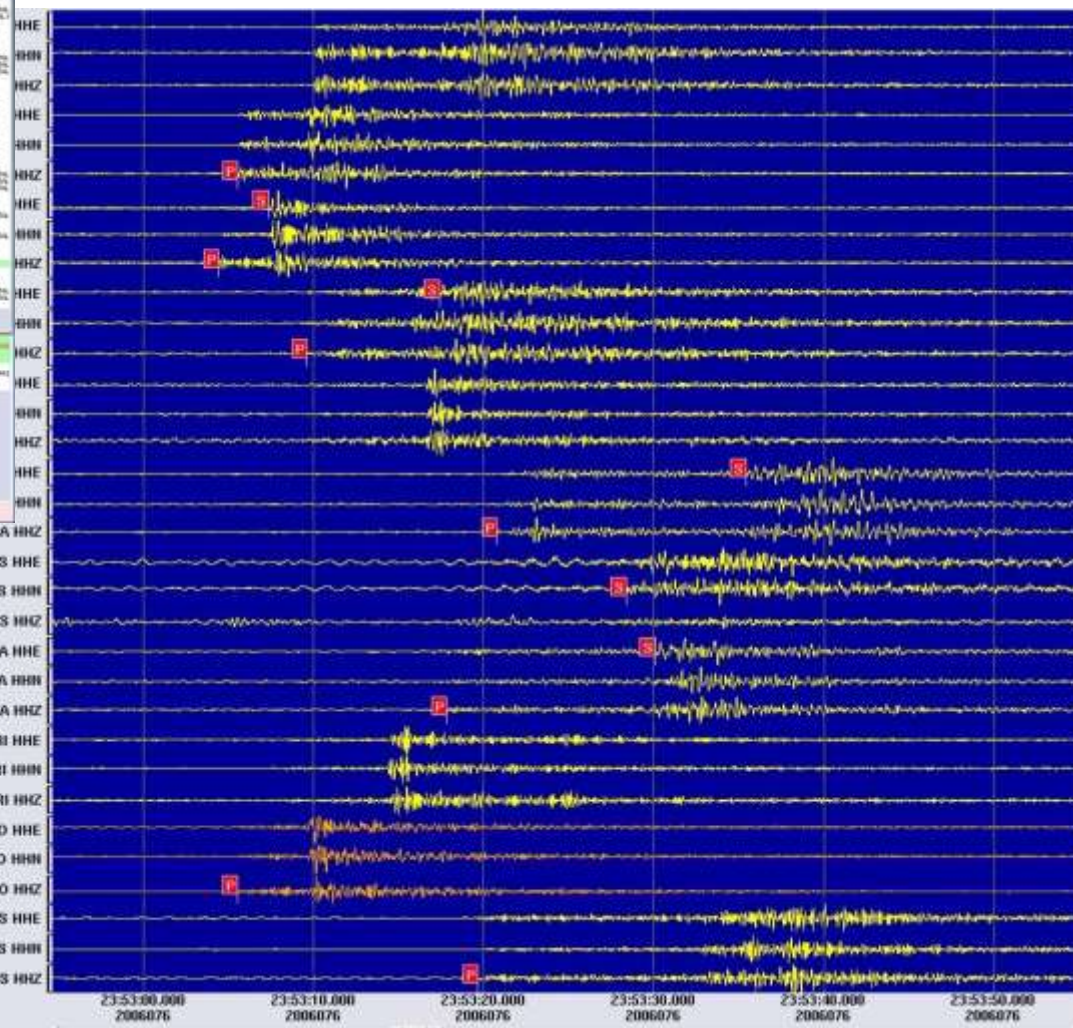
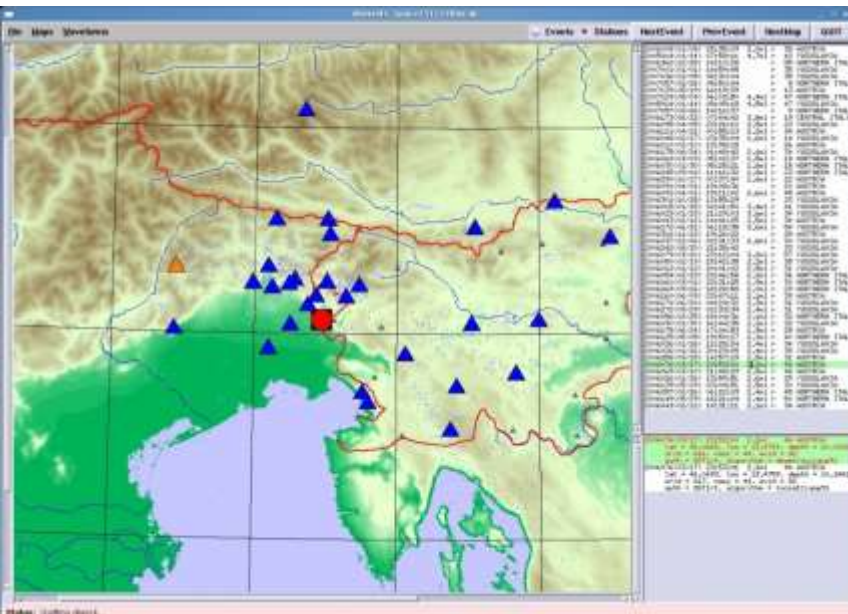


- The epicenters of around 30,000 earthquakes are now reported annually by the International Seismological Center. The geographical distribution of world seismicity dramatically illustrates the tectonically active regions of the Earth.
- The seismicity map is important evidence in support of plate tectonic theory, and delineates the presently active plate margins. Earthquake epicenters are not uniformly distributed over the Earth's surface, but occur predominantly along rather narrow zones of **interplate** seismic activity



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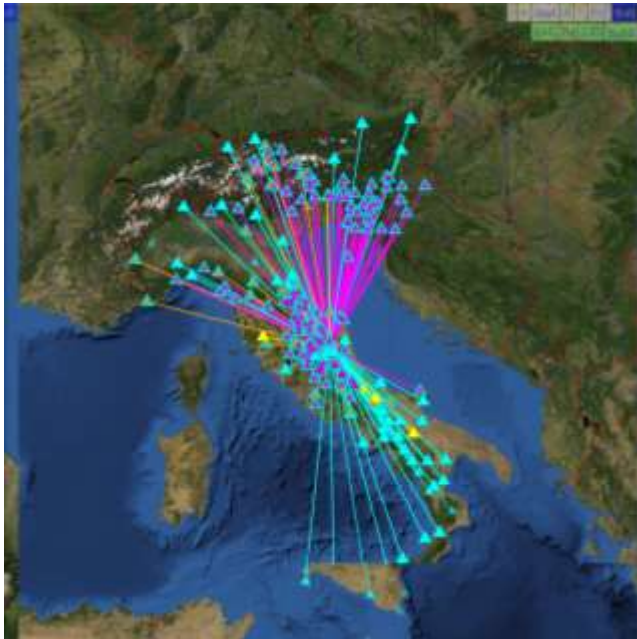




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Monitoraggio sismico





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METODO DI GEIGER Minimi quadrati

(x_i, y_i, z_i) coordinate i-esima stazione
 τ_i tempo di arrivo osservato

Assumendo un'ipocentro (x, y, z) ed un tempo origine t , sia t_i il tempo di arrivo calcolato.
Nell'ipotesi di un residuo temporale:

$$R_i \equiv \tau_i - t_i$$

Piccolo possiamo espanderlo in serie di Taylor

$$R_i = 1dt + \frac{\delta t_i}{\delta x} dx + \frac{\delta t_i}{\delta y} dy + \frac{\delta t_i}{\delta z} dz + \varepsilon_i$$

errore





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Con un dato **modello crostale** possiamo calcolare sia il tempo di arrivo che le derivate.
Il vettore di correzione ipocentrale

$$(dt, dx, dy, dz)$$

Può essere ottenuto con il metodo dei minimi quadrati, imponendo

$$\sum_i R_i^2 = \min$$

Questo porta a risolvere un sistema di equazioni normali.



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Per ottenere l'equazione normale k-esima bisogna moltiplicare ogni equazione per il coefficiente della variabile k-esima e sommare le equazioni.

In forma simbolica:

$$\begin{aligned}[1 \cdot 1]dt + [1 \cdot a]dx + [1 \cdot b]dy + [1 \cdot c]dz &= [1 \cdot R] \\ [a \cdot 1]dt + [a \cdot a]dx + [a \cdot b]dy + [a \cdot c]dz &= [a \cdot R] \\ [b \cdot 1]dt + [b \cdot b]dx + [b \cdot c]dy + [b \cdot d]dz &= [b \cdot R] \\ [c \cdot 1]dt + [c \cdot a]dx + [c \cdot b]dy + [c \cdot c]dz &= [c \cdot R]\end{aligned}$$

Ove:

$$a \equiv \frac{\partial t_i}{\partial x}$$

$$b \equiv \frac{\partial t_i}{\partial y}$$

$$c \equiv \frac{\partial t_i}{\partial z}$$

e

$$[1 \cdot 1] = \sum_{i=1}^n 1 \cdot 1 = n$$

$$[a \cdot b] = \sum_{i=1}^n a_i b_i$$

$$[1 \cdot a] = \sum_{i=1}^n a_i$$



Il Sistema da risolvere sarà:

$$ndt + \sum a_i dx + \sum b_i dy + \sum c_i dz = \sum R_i$$

$$\sum a_i dt + \sum a_i^2 dx + \sum a_i b_i dy + \sum a_i c_i dz = \sum a_i R_i$$

$$\sum b_i dt + \sum a_i b_i dx + \sum b_i b_i dy + \sum b_i c_i dz = \sum b_i R_i$$

$$\sum c_i dt + \sum a_i c dx + \sum a_i c_i dy + \sum c_i c_i dz = \sum c_i R_i$$

Questo sistema può essere risolto con il solito Sistema della matrice inversa.

Il nuovo tempo origine ed epicentro

$$t + dt, (x + dx, y + dy, z + dz)$$

Vengono usati per la nuova iterazione finchè non viene soddisfatto un criterio di arresto.



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Il problema può essere impostato schematicamente usando le matrici.
Si vuole minimizzare la somma dei quadrati dei residui R_i alle varie stazioni.
Possiamo esprimere la somma come:

$$f = \sum_{i=1}^n R_i^2 = R_i R_i = \underline{R^T R}$$

Il vettore \underline{R} è dato da:

$$\underline{R} = \underline{A} \underline{x}$$



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$$\underline{R} = \underline{A} \underline{x}$$

dove \underline{A} è la matrice delle derivate parziali dei tempi di arrivo t_i alle varie stazioni rispetto ai parametri ipocentrali e tempo origine (indicati con a, b c precedentemente);

\underline{x} è invece il vettore di correzione ipocentrale (dt, dx, dy, dz)

Il Sistema di equazioni normali si può scrivere come:

$$\underline{A}^T \underline{A} \underline{x} = \underline{A}^T \underline{R}$$

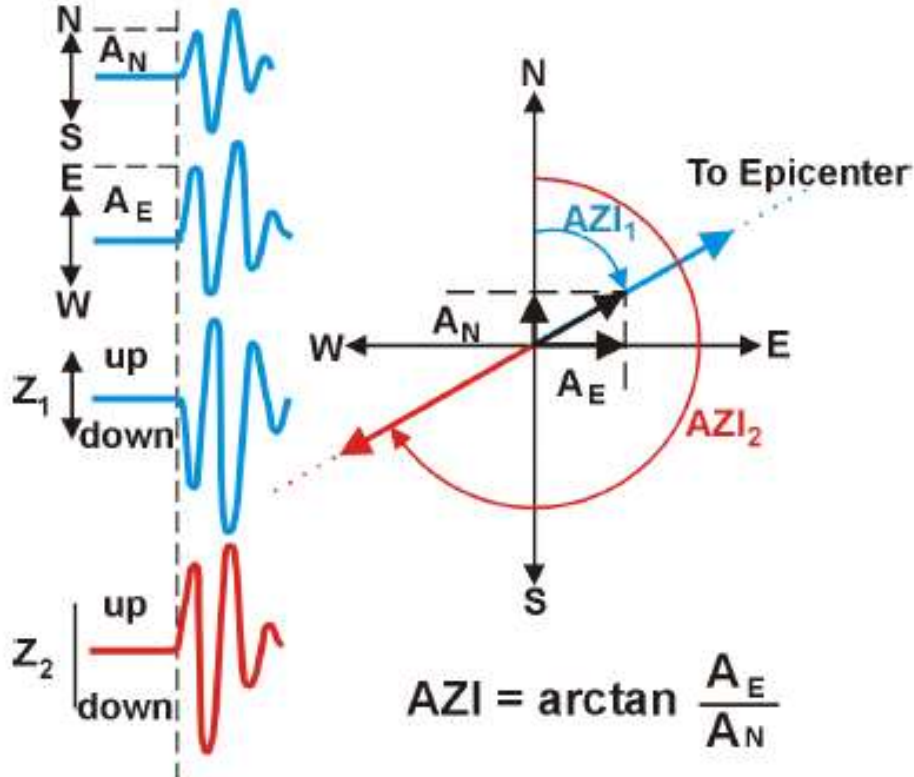
Per cui la soluzione è data da

$$\underline{x} = \left(\underline{A}^T \underline{A} \right)^{-1} \underline{A}^T \underline{R}$$

Siccome il Sistema non è lineare la procedura va iterata.

METODO DELLA STAZIONE SINGOLA

E' possibile localizzare (o avere una stima della localizzazione) un terremoto utilizzando una sola stazione a tre componenti.

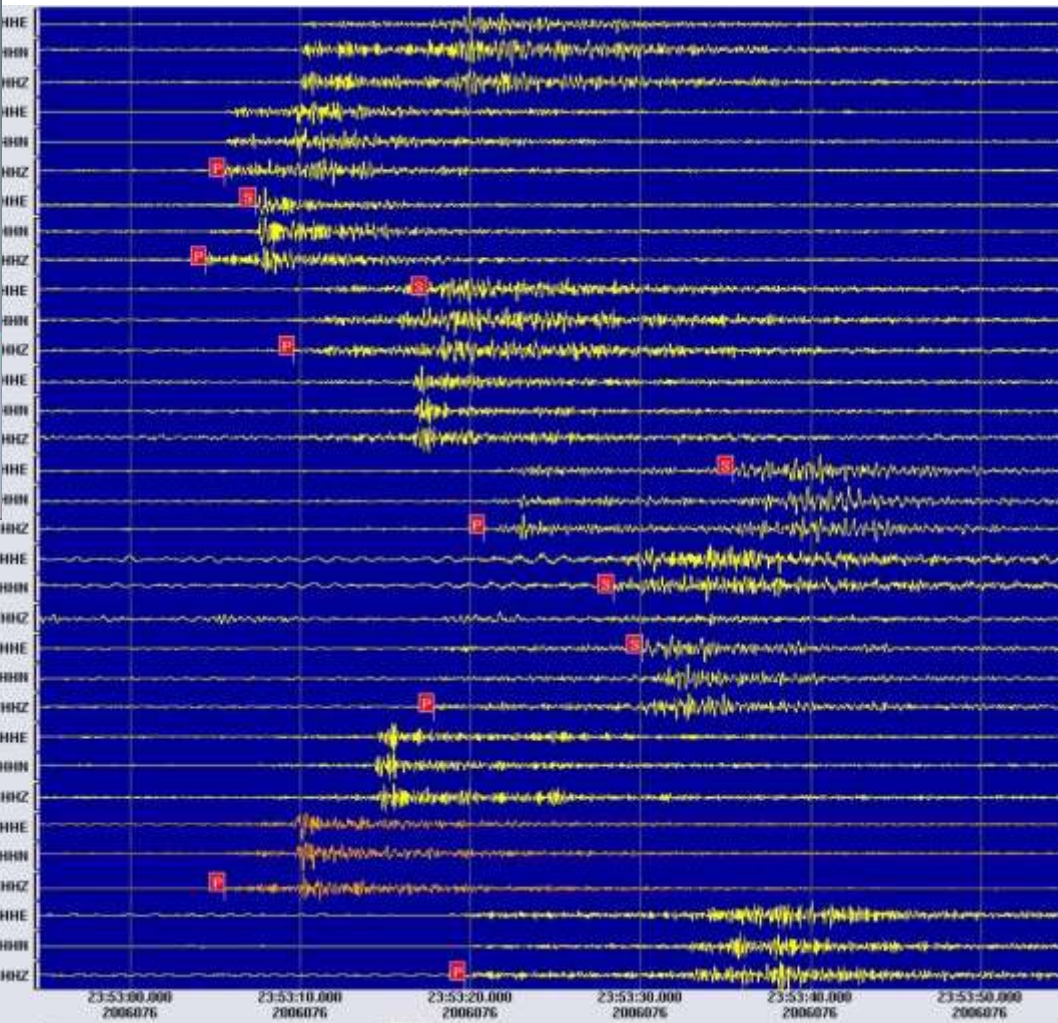
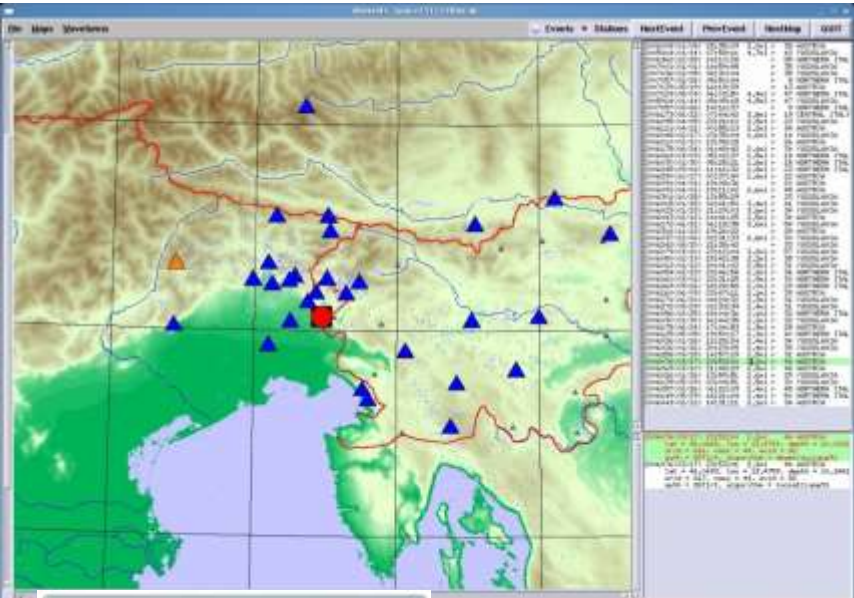


Con la componente radiale delle P si determina la direzione di provenienza, con la polarità delle P sulla verticale si determina il verso di provenienza. Il primo moto sulla verticale verso l'alto per definizione corrisponde ad un arrivo compressivo (moto in direzione uscente dalla sorgente)



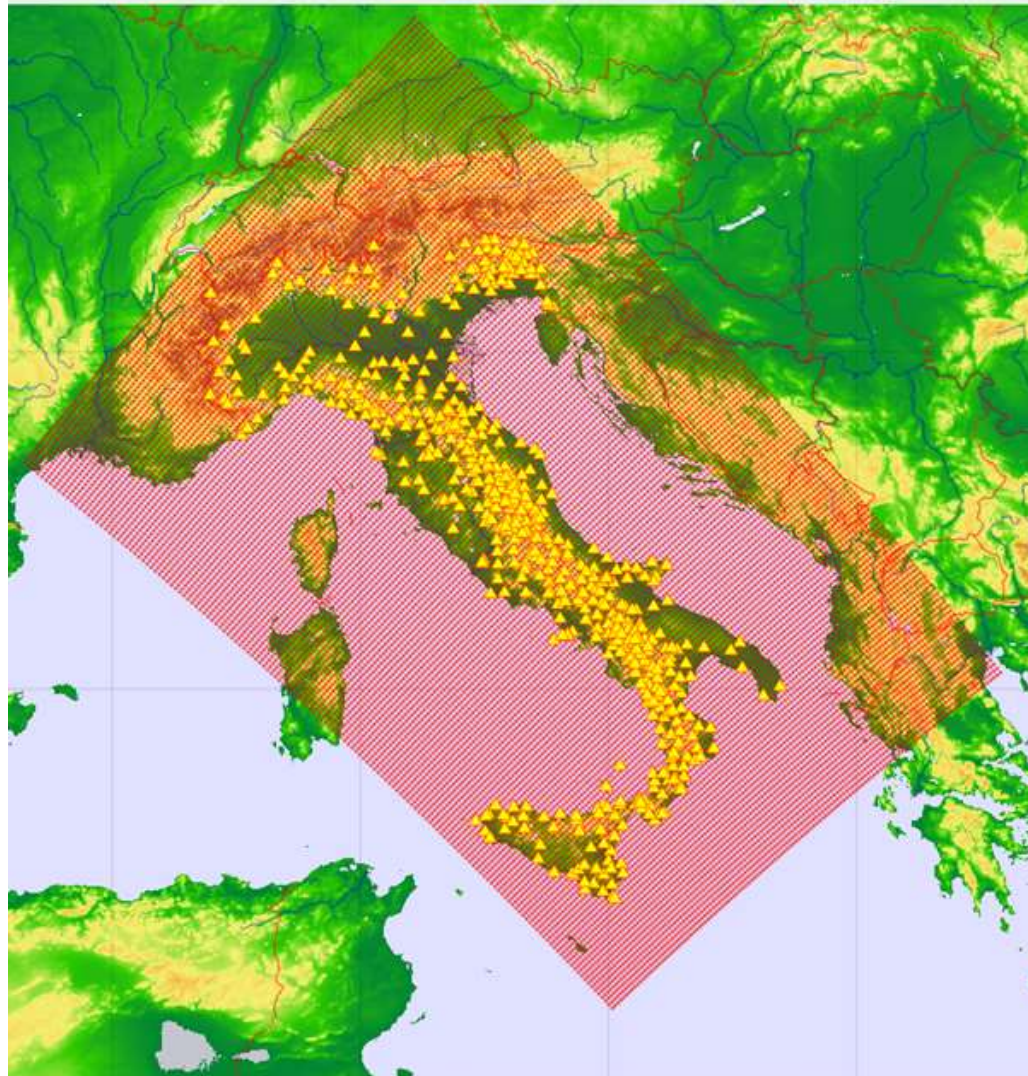
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Localizzazione su griglia





Trigger

- **A soglia**: il trigger viene attivato quando l'ampiezza del segnale supera una predefinita soglia
- **RMS** (root-mean-square): viene considerata l'ampiezza del segnale in una breve finestra temporale al posto dell'ampiezza istantanea.
- **STA/LTA**: viene considerato il rapporto della media del segnale relativo ad una finestra temporale corta (STA, short time average) rispetto ad una finestra temporale lunga (LTA, long time average).



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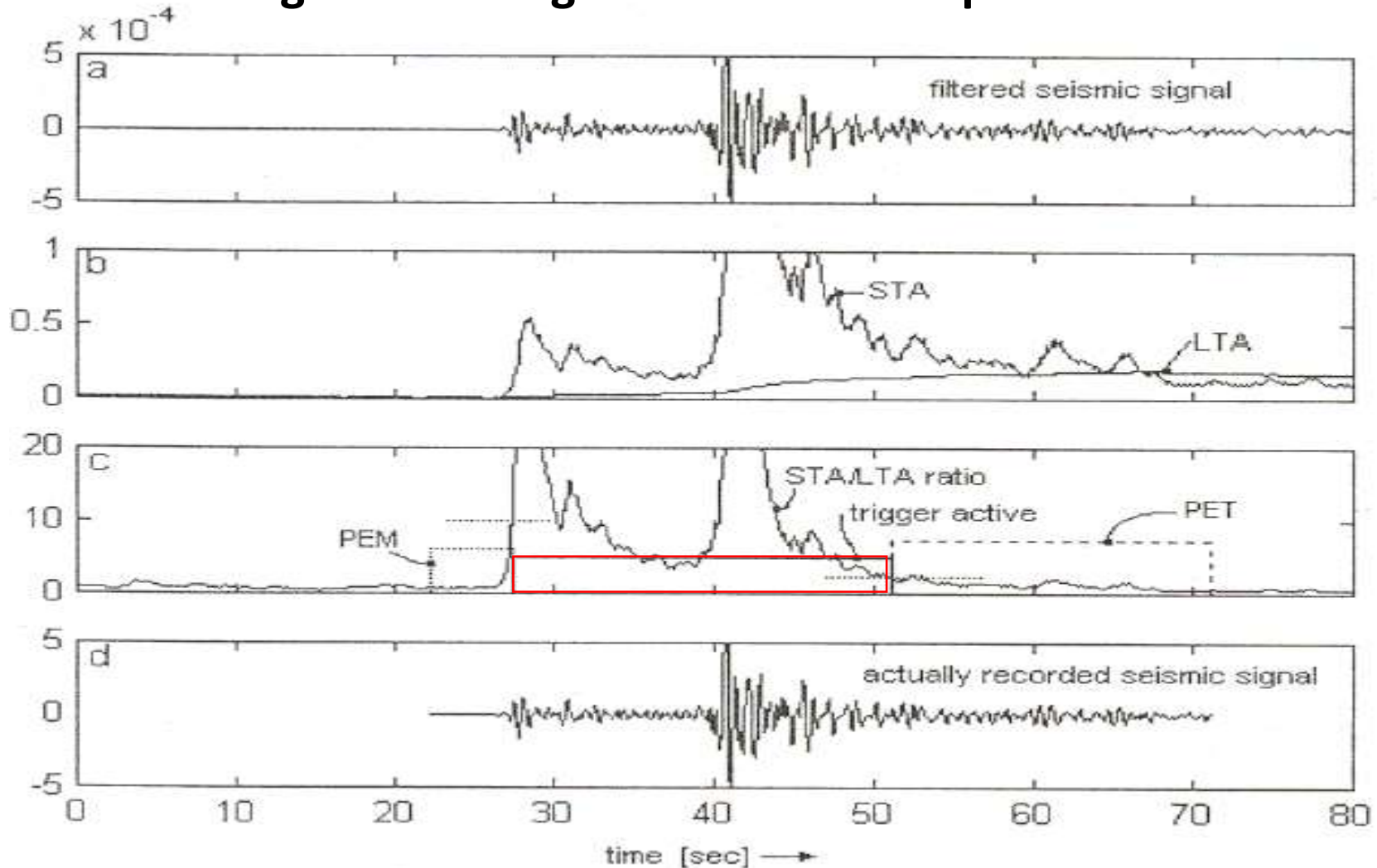
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STA – short time average

PEM – Pre-event time

LTA – long time average

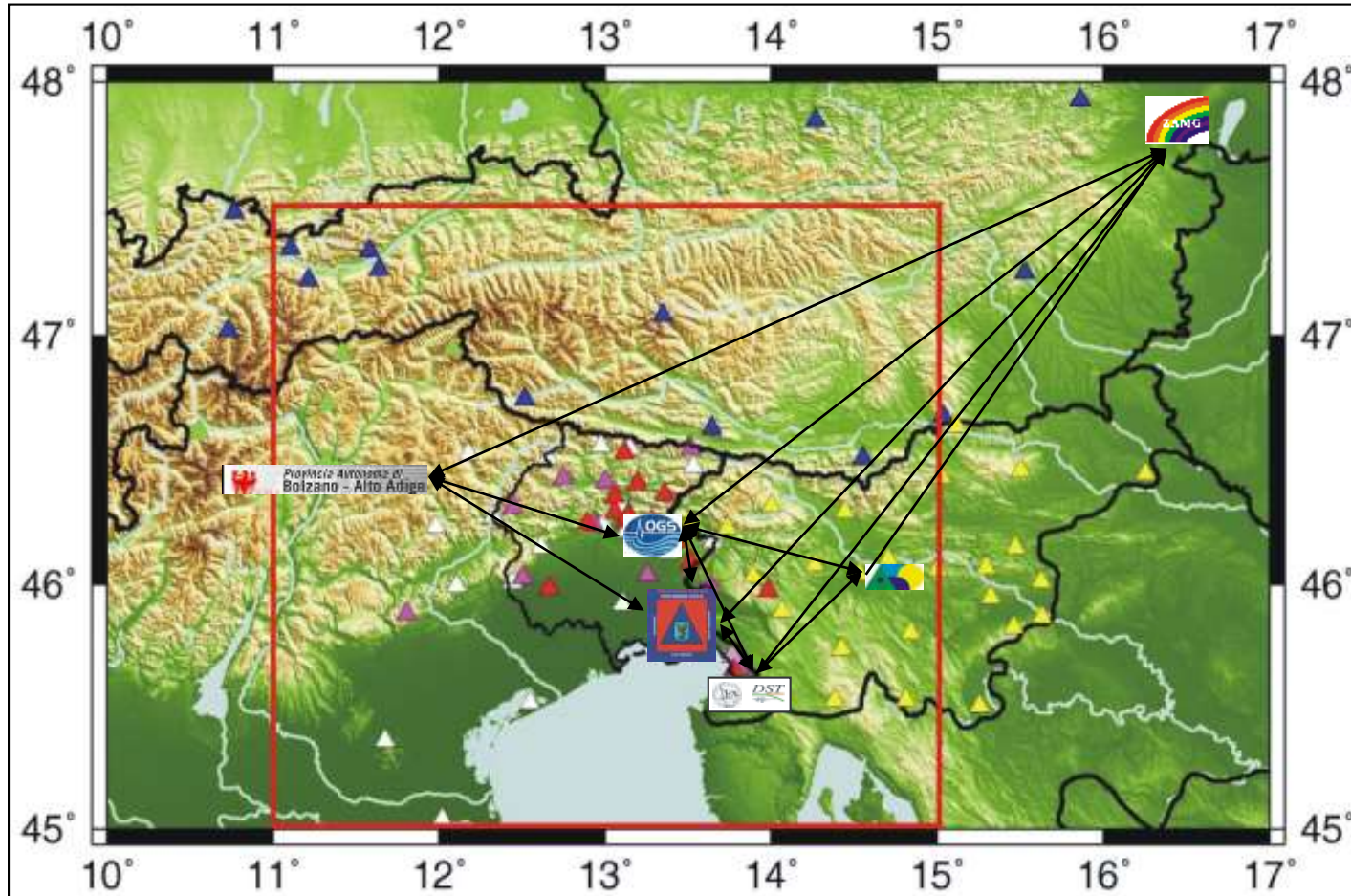
PET - post event time








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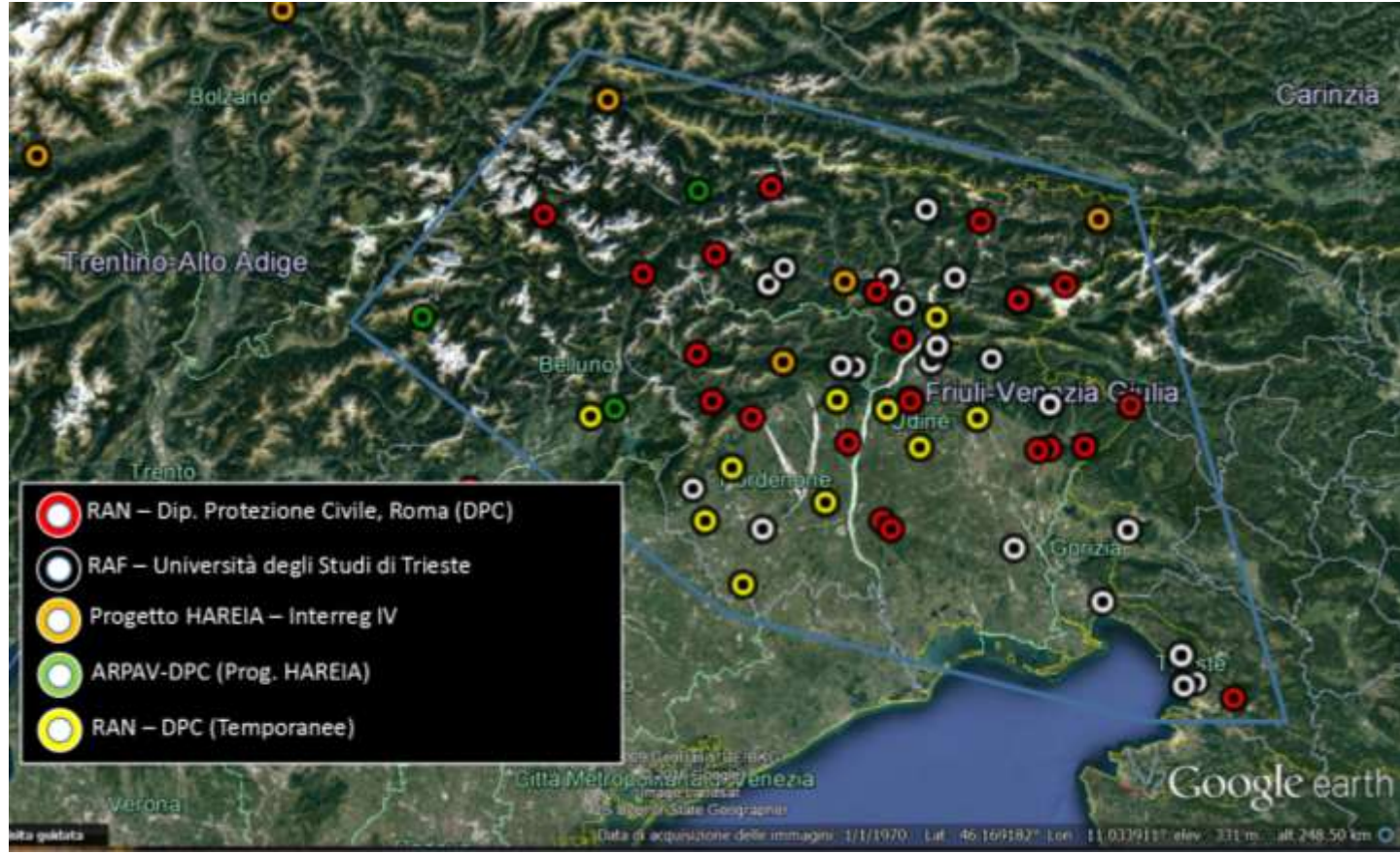


- | | | |
|--|--|---|
|  Friuli Venezia Giulia Accelerometric Network (DST) |  NE Italy Broadband Network (DST/OGS) |  Friuli Venezia Giulia and Veneto Seismometric Network (OGS) |
|  Seismic Network of Republic of Slovenia (ARSO) |  Seismic Network of Austria (ZAMG) | |



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>500 stations

RAN includes >200 stations of the historical ENEL network of the '70s, located inside electric transformer substations.



Kinematics (Etna)



>300 new stations are realized in free field, mostly on land owned by municipality close to urban areas



Syscom



Earthquake: introduction

- Most of the earthquakes which shake the Earth each year are so weak that they are only registered by sensitive seismographs, but some are strong enough to have serious, even catastrophic, consequences for mankind and the environment.
- About **90%** of all earthquakes result from **tectonic** events, primarily movements on faults. The remaining **10%** are related to **volcanism**, collapse of subterranean cavities, or man-made effects.
- Our understanding of the processes that lead to earthquakes derives to a large extent from observations of seismic events on the San Andreas fault in California.
- The average relative motion of the plates adjacent to the San Andreas fault is about 5 cm yr^{-1} , with the block to the west of the fault moving northward.
- On the fault-plane itself, this motion is not continuous but takes place spasmodically. According to modern plate tectonic theory this extensively studied fault system is a transform fault. This is a rather special type, so it cannot be assumed that the observations related to the San Andreas fault are applicable without reservation to all other faults. However, the elastic rebound model, proposed by H. F. Reid after the 1906 San Francisco quake, is a useful guide to how an earthquake may occur.



SORGENTI SISMICHE

NATURALI

TERREMOTI TETTONICI
TERREMOTI VULCANICI
IMPLOSIONI, COLLASSI
MICROSISMI OCEANICI

ARTIFICIALI

CONTROLLATE

ESPLOSIONI
RUMORE CULTURALE

INDOTTE

TERREMOTI INDOTTI DA DIGHE
TREMORI DI MINIERA
TERREMOTI INDOTTI DA
INIEZIONE DI FLUIDI



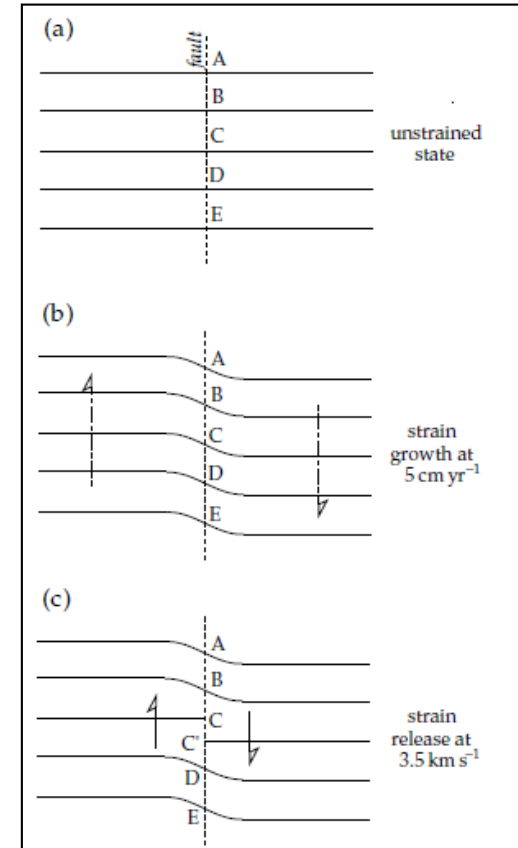
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DEFINIZIONE DI TERREMOTO TETTONICO

- 1 Il terremoto tettonico è una frattura delle rocce terrestri lungo una superficie (piana).
- 2 La frattura è dovuta agli sforzi (di taglio), che si accumulano durante la deformazione tettonica, e porta ad un rilascio totale (o parziale) degli sforzi sull'area fratturata.
- 3 La frattura inizia su un'area piccola e si propaga successivamente con velocità non maggiore della velocità delle onde longitudinali (principio di causalità).
- 4 La frattura di un terremoto tettonico è una frattura di taglio - la componente normale è trascurabile.
- 5 Il materiale nei dintorni della superficie fratturata resta essenzialmente elastico.

- The model is illustrated in Figure by the changes to five parallel lines, drawn normal to the trace of the fault in the unstrained state and intersecting it at the points A–E.
- Strain due to relative motion of the blocks adjacent to the fault accumulates over several years. Far from the trace of the fault the five lines remain straight and parallel, but close to it they are bent.
- When the breaking point of the crustal rocks at C is exceeded, rupture occurs and there is a violent displacement on the fault-plane. The relative displacement that has been taking place progressively between the adjacent plates during years or decades is achieved on the fault-plane in a few seconds.
- The strained rocks adjacent to the fault “rebound” suddenly. The accumulated strain energy is released with the seismic speed of the ruptured rocks, which is several kilometers per second.



- The segments BC and C'D undergo compression, while CD and BC' experience dilatation. The points A and E do not move; the stored strain energy at these points is not released. The entire length of the fault-plane is not displaced, only the region in which the breaking point has been exceeded. The greater the length of the fault-plane that is activated, the larger is the ensuing earthquake.



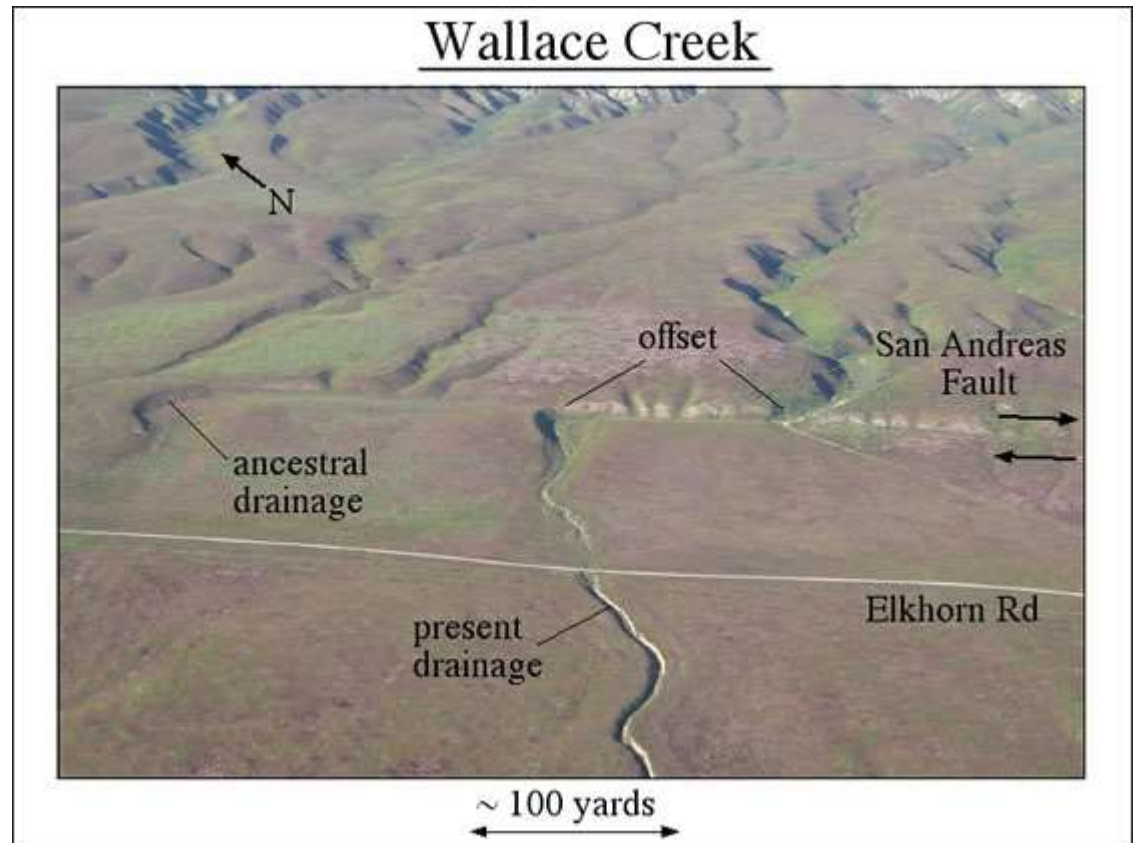
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- The occurrence of a large earthquake is not necessarily as abrupt as described in the preceding paragraph although it can be very sudden. In 1976 a major earthquake with magnitude 7.8 struck a heavily populated area of northern China near the city of Tangshan.
- Although there were known faults in the area, they had long been seismically inactive, and the large earthquake struck without warning. It completely devastated the industrial region and caused an estimated 243,000 fatalities.
- However, in many instances the accumulating strain is partially released locally as small earthquakes, or foreshocks. This is an indicator that strain energy is building up to the rupture level and is sometimes a premonition that a larger earthquake is imminent.
- When an earthquake occurs, most of the stored energy is released in the main shock. However, for weeks or months after a large earthquake there may be numerous lesser shocks, known as aftershocks, some of which can be comparable in size to the main earthquake. Structures weakened by the main event often collapse in large aftershocks, which can cause physical damage as severe as the main shock. The death toll from aftershocks is likely to be less, because people have evacuated damaged structures.



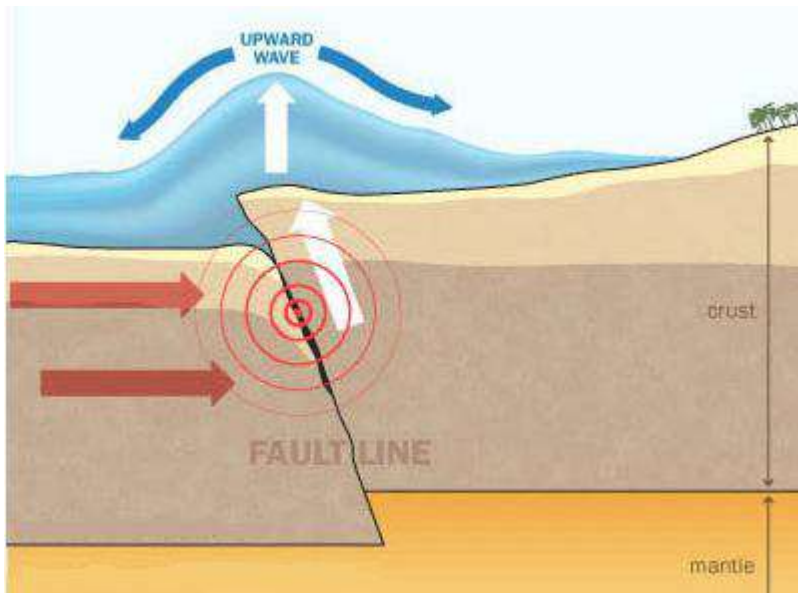
San Andreas Fault





Earthquake: elastic rebound theory

There are three different causes types of earthquakes: **tectonic**, **volcanic**, and **explosion**. The type of earthquake depends on the region where it occurs and the geological make-up of that region. The most common are tectonic earthquakes. These occur when rocks in the Earth's crust break due to geological forces created by movement of tectonic plates. Another type, volcanic earthquakes occur in conjunction with volcanic activity. Collapse earthquakes are small earthquakes in underground caverns and mines, and explosion earthquakes result from the explosion of nuclear and chemical devices.



Volcanic earthquake



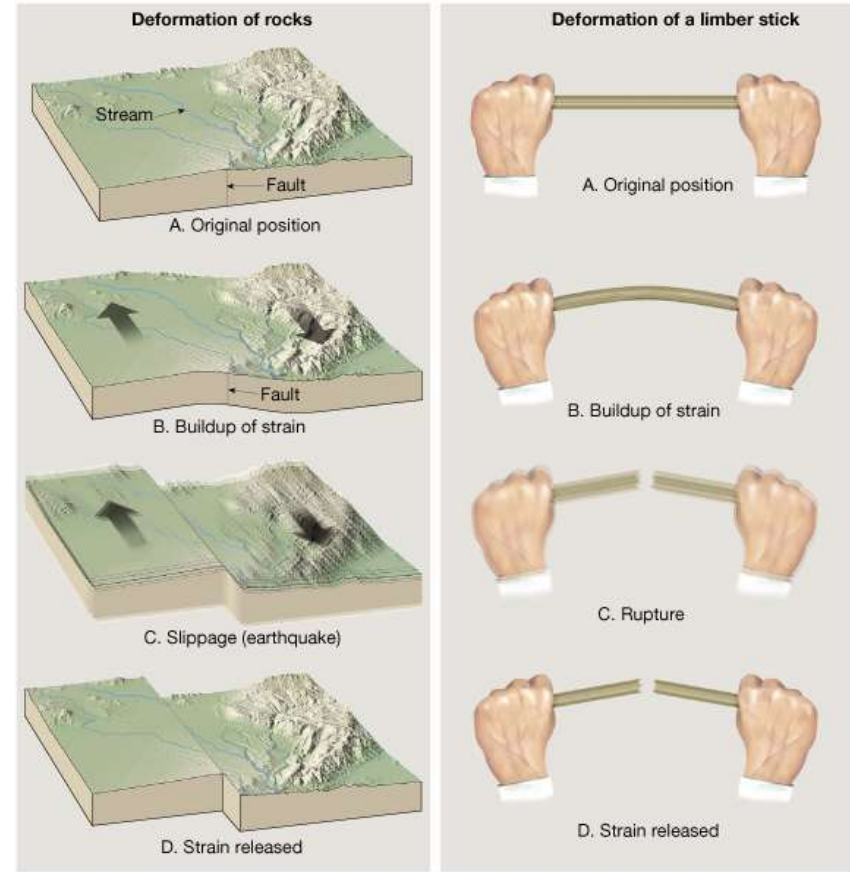
Explosion earthquake



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Tectonic earthquakes are explained by the so-called elastic rebound theory, formulated by the American geologist Harry Fielding Reid after the San Andreas Fault ruptured in 1906, generating the great San Francisco earthquake.

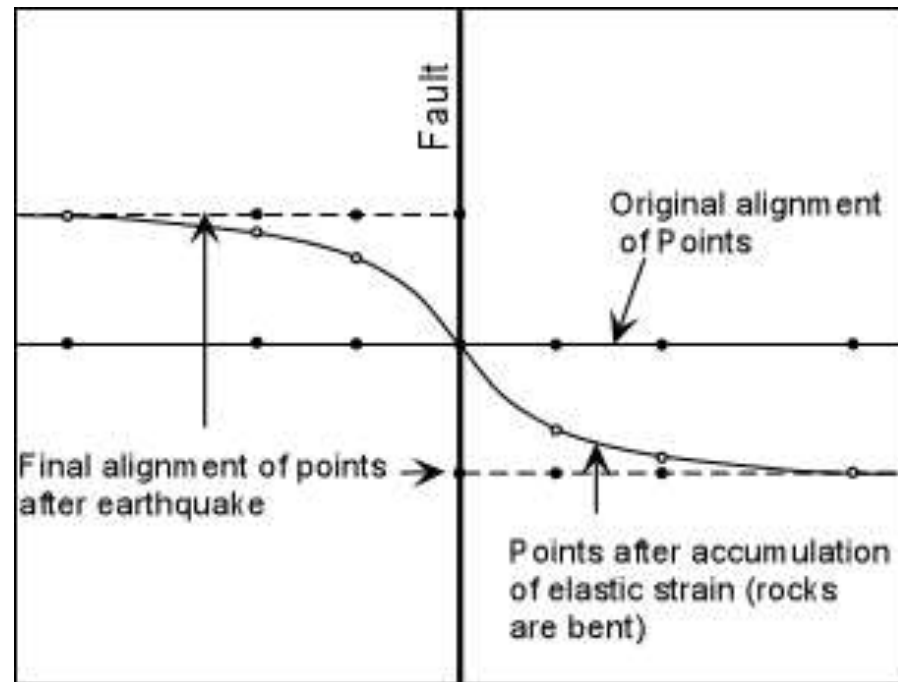
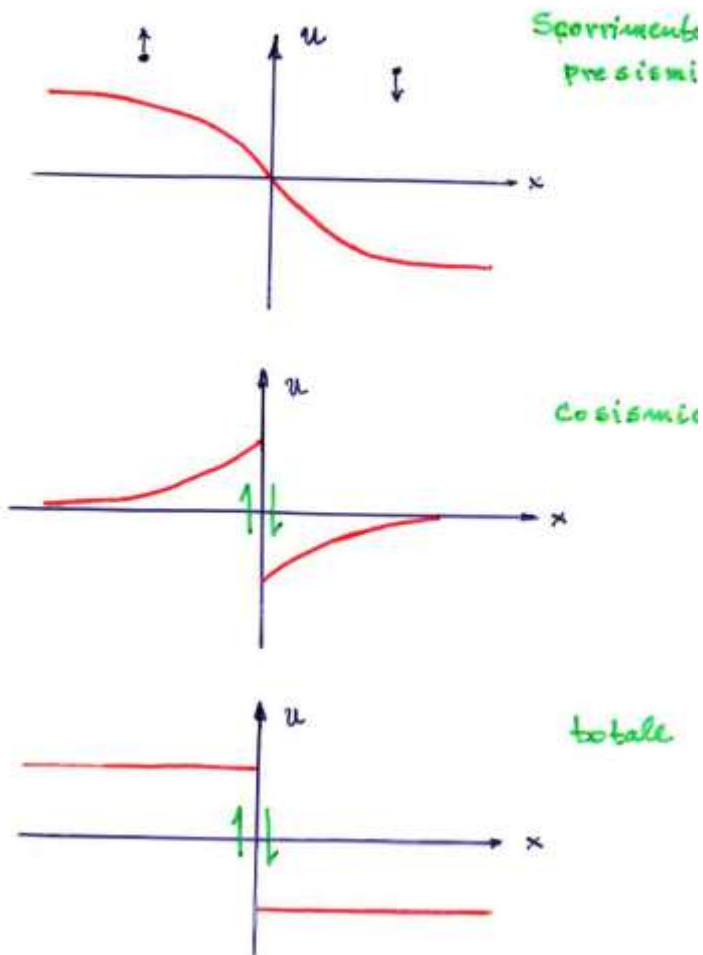




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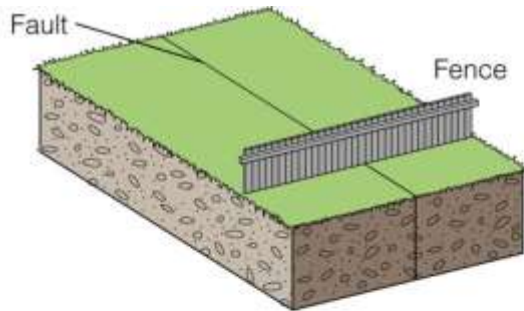
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ELASTIC REBOUND THEORY Reid

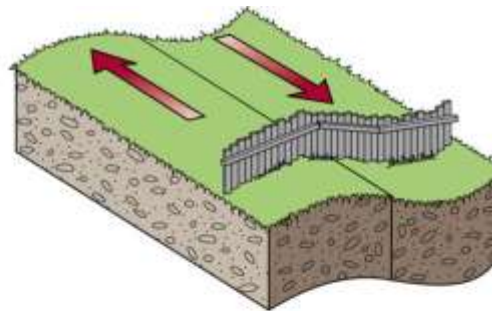




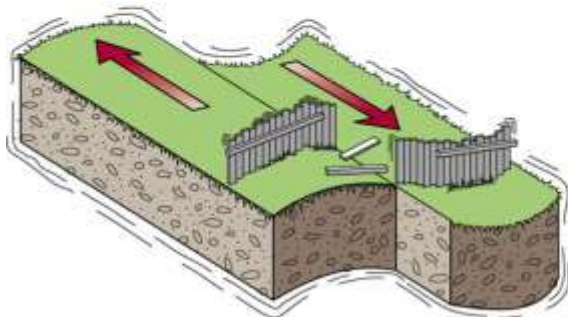
Elastic Rebound Theory



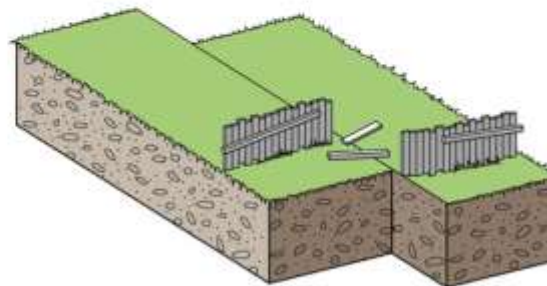
(a) Original position



(b) Deformation



(c) Rupture and release of energy



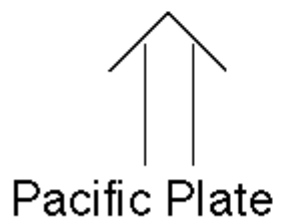
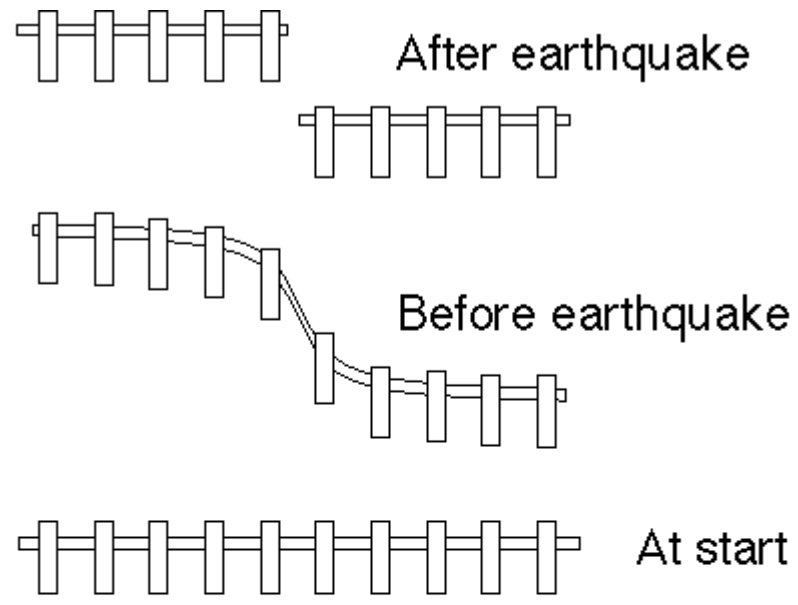
(d) Rocks rebound to original undeformed shape

1. Stress on a fault slowly accumulates
2. During an earthquake, stress on the fault is released
3. After an earthquake, stress begins to re-accumulate



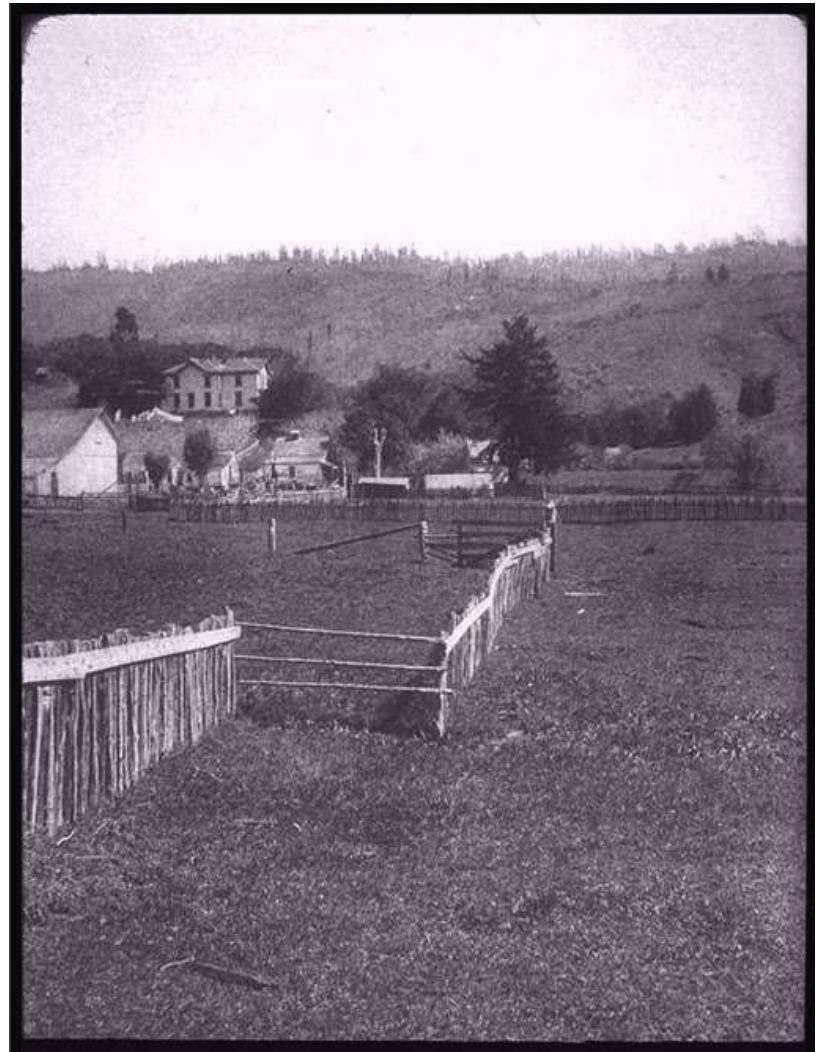
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San Andreas fault

North American Plate





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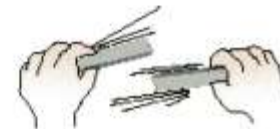
Giovanni Costa



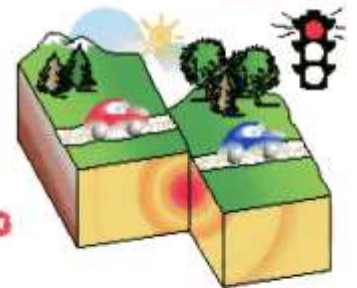
posizione originale:
SENZA DEFORMAZIONE



accumulo di energia:
PROGRESSIVA
DEFORMAZIONE



rottura con rilascio di
energia: **TERREMOTO**
E SPOSTAMENTO
PERMANENTE



La faglia



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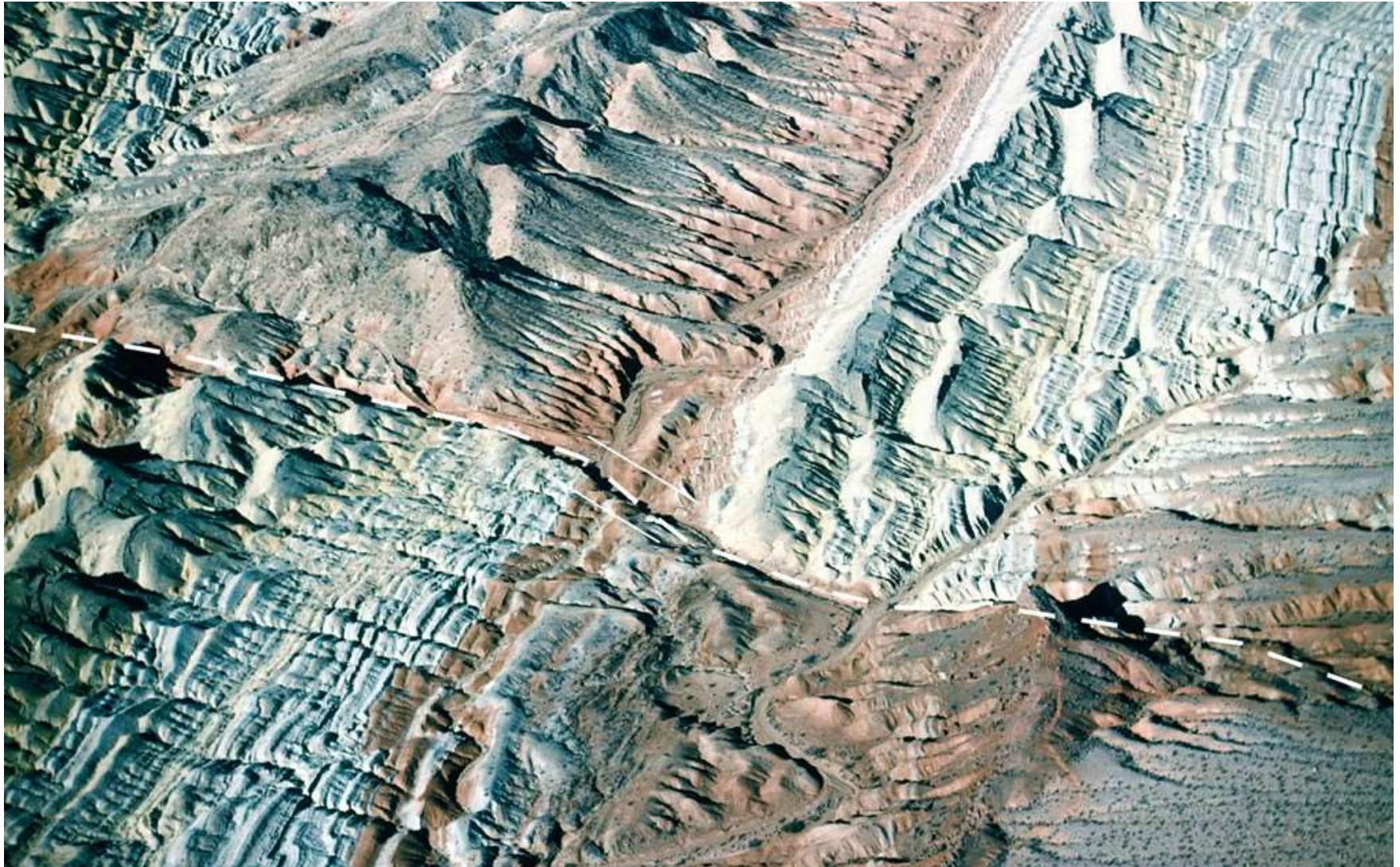
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Traccia di faglia (spostamento orizzontale)



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Lo spostamento orizzontale del terreno durante un terremoto del 1940 nella Imperial Valley, California, ha modificato la disposizione regolare degli alberi degli agrumeti (località Calexico).



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Monte Vettore

Terremoto Amatrice 24 agosto 2016



Fisica Terrestre 2022-2023

Giovanni Costa

Terremoto Amatrice 24 agosto 2016





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Terremoto Amatrice 24 agosto 2016



Esempi di nastri bianchi di neoformazione alla base del piano di faglia in roccia del Monte Vettore.



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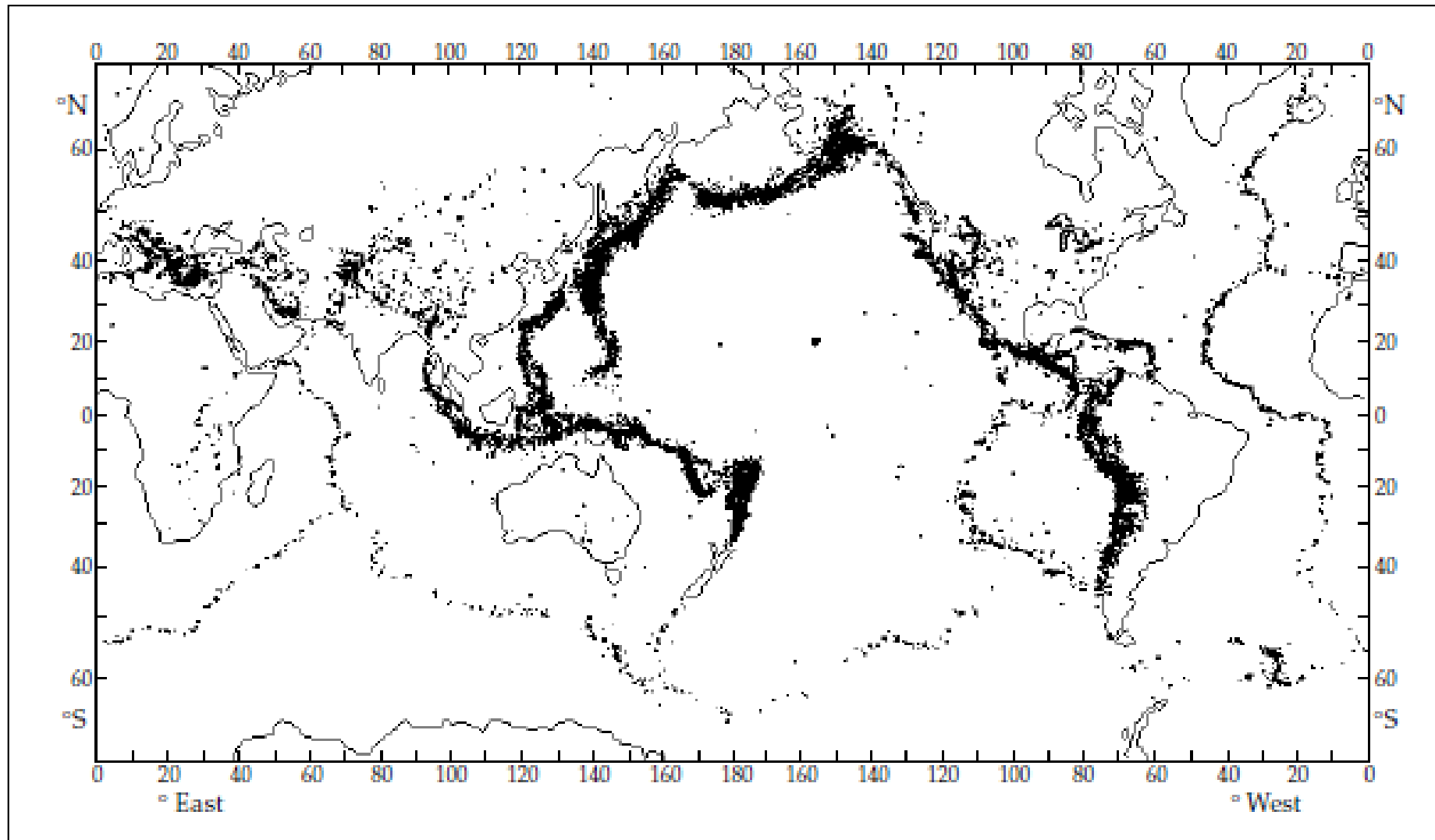
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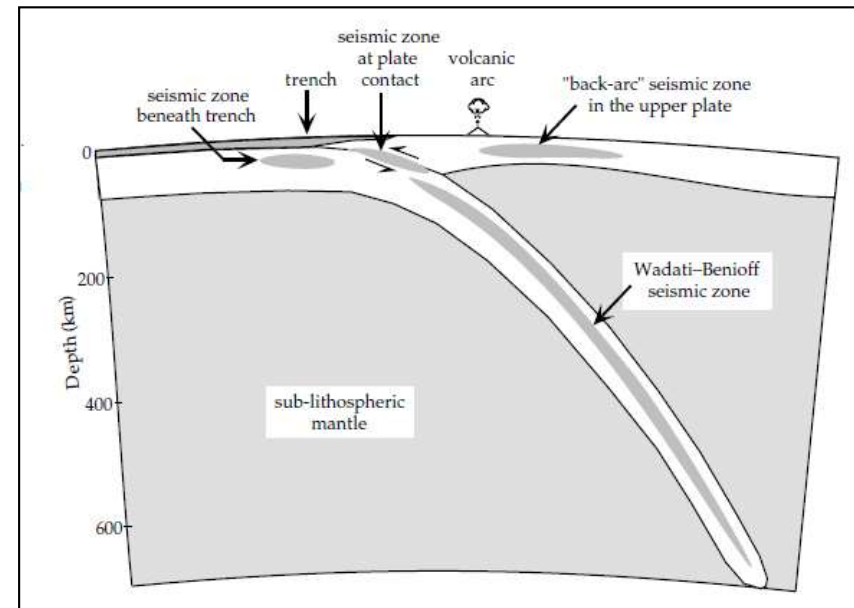




Earthquake: seismicity

- The *circum-Pacific zone*, in which about **75–80%** of the annual release of seismic energy takes place, forms a girdle that encompasses the mountain ranges on the west coast of the Americas and the island arcs along the east coast of Asia and Australasia.
- The *Mediterranean-transasiatic zone* responsible for about **15–20%** of the annual seismic energy release, begins at the Azores triple junction in the Atlantic Ocean and extends along the Azores–Gibraltar ridge; after passing through North Africa it makes a loop through the Italian peninsula, the Alps and the Dinarides; it then runs through Turkey, Iran, the Himalayan mountain chain and the island arcs of southeast Asia, where it terminates at the circum-Pacific zone.
- The system of *oceanic ridges and rises* forms the third most active zone of seismicity, with about **3–7%** of the annually released seismic energy. In addition to their seismicity, each of these zones is also characterized by active volcanism.
- The remainder of the Earth is considered to be *aseismic*
- However, no region of the Earth can be regarded as completely earthquake-free. About **1%** of the global seismicity is due to *intraplate* earthquakes, which occur remote from the major seismic zones. These are not necessarily insignificant: some very large and damaging earthquakes (e.g. the New Madrid, Missouri, earthquakes of 1811 and 1812 in the Mississippi river valley) have been of the intraplate variety.

- Earthquakes can also be classified according to their focal depths. Earthquakes with shallow focal depths, less than **70 km**, occur in all the seismically active zones; only shallow earthquakes occur on the oceanic ridge systems.
- The largest proportion (**about 85%**) of the annual release of seismic energy is liberated in **shallow-focus** earthquakes.
- The remainder is set free by earthquakes with intermediate focal depths of **70–300 km** (about **12%**) and by earthquakes with deep focal depths greater than **300 km** (about **3%**).
- These occur only in the circum-Pacific and Mediterranean-transasiatic seismic zones, and accompany the process of plate subduction.
- The distributions of epicentral locations and focal depths of intermediate and deep earthquakes give important evidence for the processes at a subduction zone.
- When the earthquake foci along a subduction zone are projected onto a cross-section normal to the strike of the plate margin, they are seen to define a zone of seismicity about 30–40 km thick in the upper part of the 80–100 km thick subducting oceanic plate, which plunges at roughly 30–60° beneath the overriding plate
- For many years the inclined seismic zone was referred to in Western literature as a Benioff zone in recognition of the Californian scientist, Hugo Benioff.

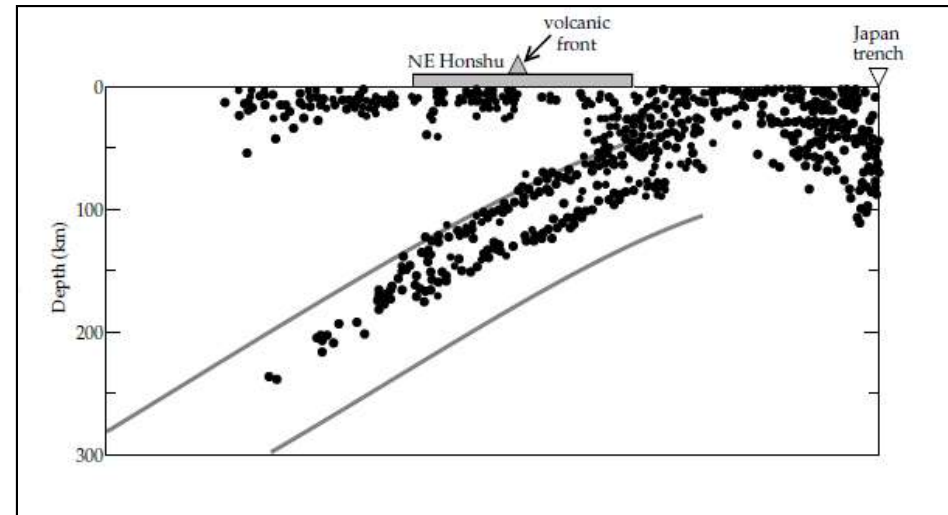




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- In the years following World War II Benioff carried out important pioneering studies that described the distribution of deep earthquakes on steeply dipping surfaces of seismicity
- Many characteristics of the occurrence of deep earthquakes had been described in the late 1920s by a Japanese seismologist, Kiyoo Wadati. He discovered that the closer the epicenters of earthquakes lay to the Asian continent, the greater were their focal depths; the deep seismicity appeared to lie on an inclined plane.

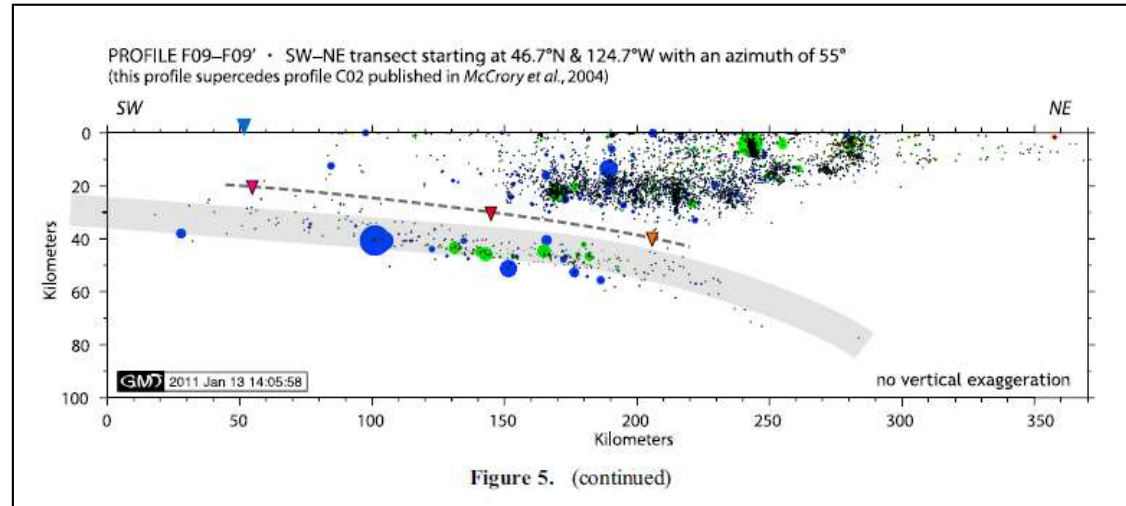
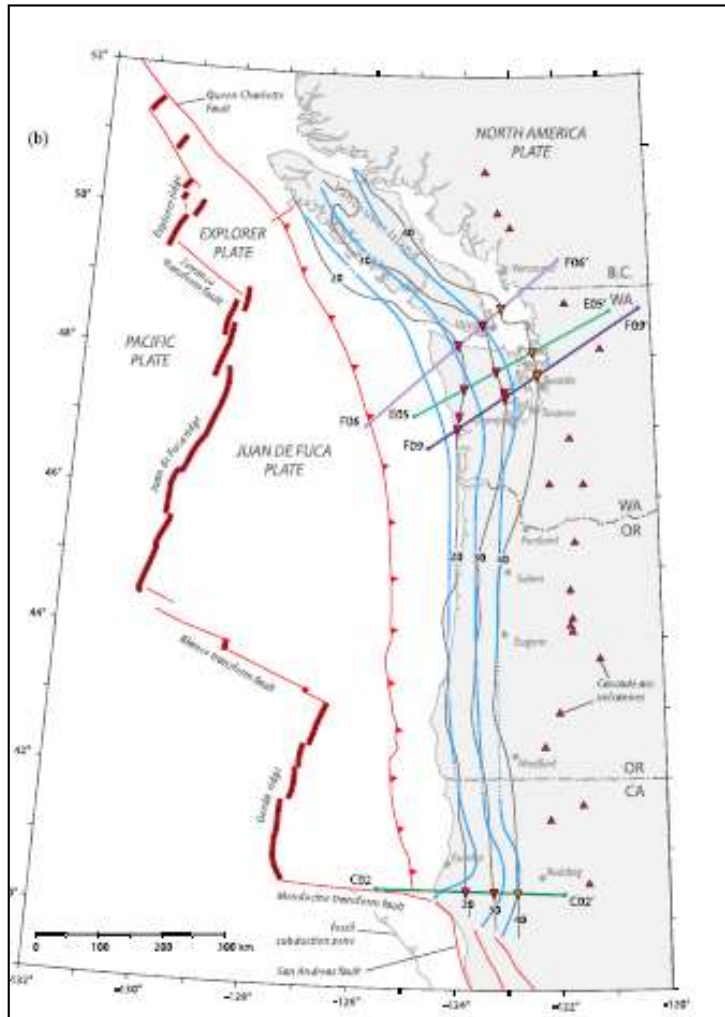


- It was Benioff, however, who in 1954 proposed as an explanation of the phenomenon that the ocean floor was being “subducted” underneath the adjacent land. This was a bold proposal well in advance of the advent of plate tectonic theory.
- Today the zone of active seismicity is called a Wadati–Benioff zone in recognition of both discoverers.



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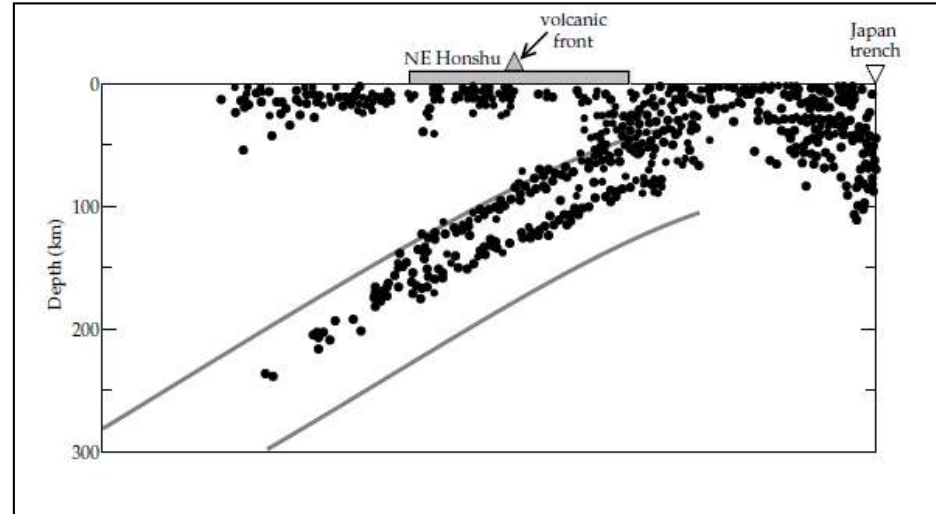
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- In three dimensions the Wadati–Benioff zone constitutes an inclined slab dipping underneath the overriding plate. It marks the location and orientation of the upper surface of the subducting plate. The dip-angle of the zone varies between about 30° and 60° , becoming steeper with increasing depth, and it can extend to depths of several hundred kilometers into the Earth. The deepest reliably located focal depths extend down to about 680 km.
- The structure of a subducting plate is not always as simple as described: Pacific plate revealed a double Wadati–Benioff zone under northeast Honshu, Japan. The seismicity at depths below 100 km defines two parallel planes about 30–40 km apart. The upper plane, identified with the top of the subducting plate, is in a state of compression; the lower plane, in the middle of the slab, is in a state of extension.
- These stress states are the result of unbending of the subducting plate, which had previously undergone sharp bending at shallow depth below the trench axis. This information is inferred from analysis of the **mechanisms** by which the earthquakes occur.



Earthquake: size

- There are two methods of describing how large an earthquake is. The **intensity** of the earthquake is a subjective parameter that is based on an assessment of visible effects.
- It therefore depends on factors other than the actual size of the earthquake. The **magnitude** of an earthquake is determined instrumentally and is a more objective measure of its size, but it says little directly about the seriousness of the ensuing effects.
- Illogically, it is usually the magnitude that is reported in news coverage of a major earthquake, whereas the intensity is a more appropriate parameter for describing the severity of its effects on mankind and the environment.



Earthquake: intensity

- Large earthquakes produce alterations to the Earth's natural surface features, or severe damage to man-made structures such as buildings, bridges and dams. Even small earthquakes can result in disproportionate damage to these edifices when inferior constructional methods or materials have been utilized. The intensity of an earthquake at a particular place is classified on the basis of the local character of the visible effects it produces.
- It depends very much on the acuity of the observer, and is in principle subjective. Yet, intensity estimates have proved to be a viable method of assessing earthquake size, including historical earthquakes.
- The first attempt to grade earthquake severity was made in the late eighteenth century by Domenico Pignataro, an Italian physician, who classified more than 1000 earthquakes that devastated the southern Italian province of Calabria in the years 1783–1786. His crude analysis classified the earthquakes according to whether they were very strong, strong, moderate or slight.
- In the mid-nineteenth century an Irish engineer, Robert Mallet, produced a list of 6831 earthquakes and plotted their estimated locations, producing the first map of the world's seismicity and establishing that earthquakes occurred in distinct zones. He also used a four-stage intensity scale to grade earthquake damage, and constructed the first isoseismal maps with lines that outlined areas with broadly equal grades of damage.



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- The Rossi–Forel intensity scale, developed in the late nineteenth century by the Italian scientist M. S. de Rossi and the Swiss scientist F. Forel, incorporated ten stages describing effects of increasing damage.
- In 1902 a Italian seismologist, G. Mercalli, proposed a still more extensive, expanded intensity scale which reclassified earthquake severity in twelve stages.
- A variation, the Modified Mercalli (MM) scale, was developed in 1931 to suit building conditions in the United States, where a later modification is in common use. The Medvedev–Sponheuer–Karnik (MSK) scale, introduced in Europe in 1964, and modified in 1981, also has twelve stages and differs from the MM scale mainly in details.
- A new European Macroseismic Scale (EMS-98) was adopted in 1998; an abridged version. The new 12-stage EMS scale is based on the MSK scale but takes into account the vulnerability of buildings to earthquake damage and incorporates more rigorous evaluation of the extent of damage to structures with different building standards.



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a) Compilata da Mercalli e modificata nel 1931

- I. Non avvertito
- II. Avvertito negli interni solo da poche persone per lo più nei piani alti delle case e da persone in posizione favorevole
- III. Avvertito negli interni ove gli oggetti pendenti oscillano. Vibrazioni come se passassero camion leggeri. E' possibile stimarne la durata. Può non essere riconosciuto come un terremoto.
- IV. Gli oggetti pendenti oscillano. Vibrazioni come se passassero camion pesanti, oppure sensazione di colpo secco come quello di una palla pesante contro un muro. Le automobili ferme oscillano. Finestre, piatti, bicchieri e porte vibrano. Al limite superiore del IV grado le pareti di legno e gli infissi si incrinano.
- V. Avvertito anche esternamente. Coloro che dormono si svegliano. I liquidi oscillano e a volte traboccano. Piccoli oggetti instabili si muovono e si ribaltano. Le porte si aprono e si chiudono. Imposte e quadri si muovono. Gli orologi a pendolo si fermano o cominciano a funzionare e cambiano periodo di oscillazione.
- VI. Avvertito da tutti. Molti impauriti corrono fuori dagli edifici. Finestre, piatti, bicchieri si rompono. Cadono i libri dagli scaffali ed i quadri dalle pareti. I mobili si muovono o si rovesciano. Le costruzioni fragili si incrinano. Piccole campane (come quelle delle chiese e delle scuole) suonano. Gli alberi ed i cespugli ondeggiano.
- VII. Difficile reggersi in equilibrio. Avvertito da quelli che sono alla guida di una automobile. I mobili si rompono. Danni moderati alle costruzioni in fango. Camini fragili si rompono alla base. Cadono intonaci, mattoni sciolti, pietre, cornicioni. Qualche incrinatura nelle costruzioni in pietra. Si formano onde nelle pozzanghere l'acqua diventa torbida con fango. Piccole frane e avvalamenti nei banchi di sabbia e ghiaia. Campane grandi suonano. Canali d'irrigazione in cemento sono danneggiati.





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- VIII. La guida delle automobili è difficile. Danni alle costruzioni in pietra; collasso parziale. Qualche danno di stucchi e di qualche muro in mattoni. Oscillazioni o crollo di camini, ciminiera, monumenti, torri, serbatoi sospesi. Gli edifici oscillano sulle fondamenta; i tramezzi cadono. Si spezzano i rami degli alberi. Il flusso e la temperatura di sorgenti e di pozzi cambiano. Si formano fratture nel terreno umido e in forti pendenze.
- IX. Panico generale. Costruzioni di fango completamente distrutte; costruzioni in pietra fortemente danneggiate, a volte un collasso totale; costruzioni in mattoni seriamente danneggiate. Gli edifici crollano o sono spostati dalle fondamenta. Forti danni ai serbatoi. I tubi sotterranei si rompono. Si formano fratture cospicue nel terreno. In aree alluvionali sono spruzzati sabbia e fango; si formano crateri di sabbia.
- X. La maggioranza delle costruzioni è distrutta. Le strutture in legno, anche se ben costruite, ed i ponti sono distrutti. Forti danni alle dighe, ai canali di irrigazione e agli argini. Grandi frane. Invasione di acqua proveniente da canali, fiumi, laghi, ecc. Sabbia e fango sono spostati orizzontalmente sulle spiagge e nelle regioni pianeggianti. Le rotaie sono leggermente incurvate.
- XI. Le rotaie sono fortemente incurvate. I tubi sotterranei sono completamente inutilizzabili.
- XII. Danno quasi totale. Grandi massi di roccia spostati. La morfologia è alterata. Oggetti lanciati in aria.



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Intensity	Description of effects
<i>I-IV light to moderate earthquakes</i>	
I	Not felt
II	Scarcely felt Felt only by a few individual people at rest in houses.
III	Weak Felt indoors by a few people. People at rest feel a swaying or light trembling.
IV	Largely observed Felt indoors by many people; outdoors by very few. A few people are awakened. Windows, doors and dishes rattle.
<i>V-VIII moderate to severe earthquakes</i>	
V	Strong Felt indoors by most, outdoors by few. Many sleeping people awake. A few are frightened. Buildings tremble throughout. Hanging objects swing considerably. Small objects are shifted. Doors and windows swing open or shut.
VI	Slightly damaging Many people are frightened and run outdoors. Some objects fall. Many houses suffer slight non-structural damage like hair-line cracks and fall of small pieces of plaster.
VII	Damaging Most people are frightened and run outdoors. Furniture is shifted and objects fall from shelves in large numbers. Many well built ordinary buildings suffer moderate damage: small cracks in walls, fall of plaster, parts of chimneys fall down; older buildings may show large cracks in walls and failure of fill-in walls.
VIII	Heavily damaging Many people find it difficult to stand. Many houses have large cracks in walls. A few well built ordinary buildings show serious failure of walls, while weak older structures may collapse.
<i>IX-XII severe to destructive earthquakes</i>	
IX	Destructive General panic. Many weak constructions collapse. Even well built ordinary buildings show very heavy damage: serious failure of walls and partial structural failure.
X	Very destructive Many ordinary well built buildings collapse.
XI	Devastating Most ordinary well built buildings collapse, even some with good earthquake resistant design are destroyed.
XII	Completely devastating Almost all buildings are destroyed.

- The scale focuses especially on the effects on people and buildings. It takes into account classifications of both the vulnerability of a structure (i.e., the materials and method of construction) and the degree of damage.



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Damage %	Rossi-Forel	Modified Mercalli	Geoffan	PRC	JMA	MSK	
0	I	I	I	I	0	I	
			II			II	
	II	II	III	II	I	III	
	III		IV				
	IV	IV	IV	IV	II	IV	
	V						V
	VI	VI	VI	VI	III	V	
	VII						VII
	10	VIII	VII	VII	VII	IV	VI
	20	IX	IX	IX	IX	V	VII
	30						
40	X	X	X	X	VI	IX	
50							XI
70	X	XI	XI	XI	VII	XI	
90							XII

Figure 1



Intensità

Misura gli effetti di un terremoto sulle persone, sugli edifici e sul territorio.

SCALA MERCALLI	
I - Strumentale	Avvertita solo dagli strumenti
II - Debole	Avvertita solo da poche persone sensibili in condizioni particolari
III - Leggera	Avvertita da poche persone
IV - Moderata	Avvertita da molte persone; tremiti di infissi e cristalli; oscillazione di oggetti sospesi
V - Piuttosto forte	Avvertita da molte persone, anche addormentate; caduta di oggetti
VI - Forte	Qualche lesione agli edifici
VII - Molto forte	Caduta di comignoli; lesione agli edifici
VIII - Distruttiva	Rovina parziale di alcuni edifici; vittime isolate
IX - Rovinosa	Rovina totale di alcuni edifici; molte vittime; crepacci nel suolo
X - Disastrosa	Crollo di parecchi edifici; numerose vittime; crepacci evidenti nel terreno
XI - Molto disastrosa	Distruzione di agglomerati urbani; moltissime vittime; crepacci; frane; maremoto
XII - Catastrofica	Danneggiamento totale; distruzione di ogni manufatto; pochi superstiti; sconvolgimento del suolo; maremoto



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Intensità

Ha Gestito il Terremoto x Ha Gestito il Terremoto - D

← → ↻ 🏠 [haerstoffemerito.it](https://www.haerstoffemerito.it) https://www.haerstoffemerito.it

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Questo sito è solo per monitorare, in tempo reale, gli effetti dei terremoti. Non è per fare previsioni né per allarmare. Questo è possibile perché il sistema è alimentato solo con dati in tempo reale. Questo sistema non è in grado di prevedere il tempo e il luogo di un terremoto. Il sistema è alimentato solo con dati in tempo reale. Questo sistema non è in grado di prevedere il tempo e il luogo di un terremoto. Il sistema è alimentato solo con dati in tempo reale. Questo sistema non è in grado di prevedere il tempo e il luogo di un terremoto.

*** campi obbligatori**

DATA E ORA DEL TERREMOTO

Seleziona dalla lista il terremoto a cui ti riferisci. Se non lo trovi seleziona "Terremoto non in lista" e specifica data e ora dell'evento.

Seleziona il terremoto dalla lista

Terremoto non in lista

DATI DI LOCALITÀ

Regione

Provincia

Comune

Frazione

Indirizzo

Numero civico

Distretto

Piano

Numero totale dei piani dell'edificio

Casa Main building

HA AVVERTITO IL TERREMOTO O OSSERVATO I SUOI EFFETTI ?

Sì

No

INFORMAZIONI PERSONALI (OPZIONALI)

Il tuo contributo al questionario sarà utile per migliorare il servizio. Per maggiori informazioni sui dati e sui servizi visitate il sito www.haerstoffemerito.it. Per maggiori informazioni sui dati e sui servizi visitate il sito www.haerstoffemerito.it. Per maggiori informazioni sui dati e sui servizi visitate il sito www.haerstoffemerito.it.

Nome

Cognome

Nome

VIA

Stato

Telefono



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Intensità





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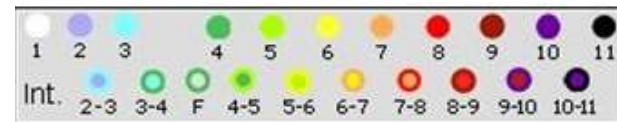
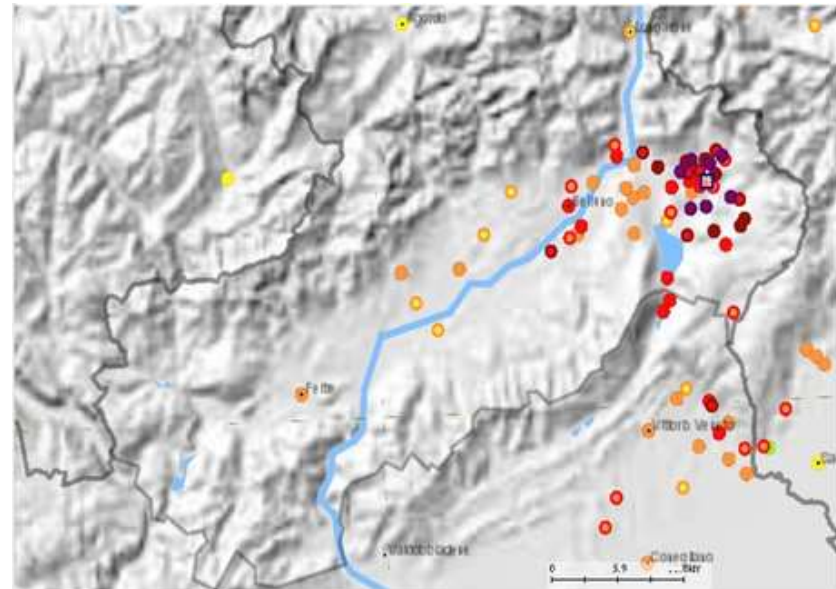
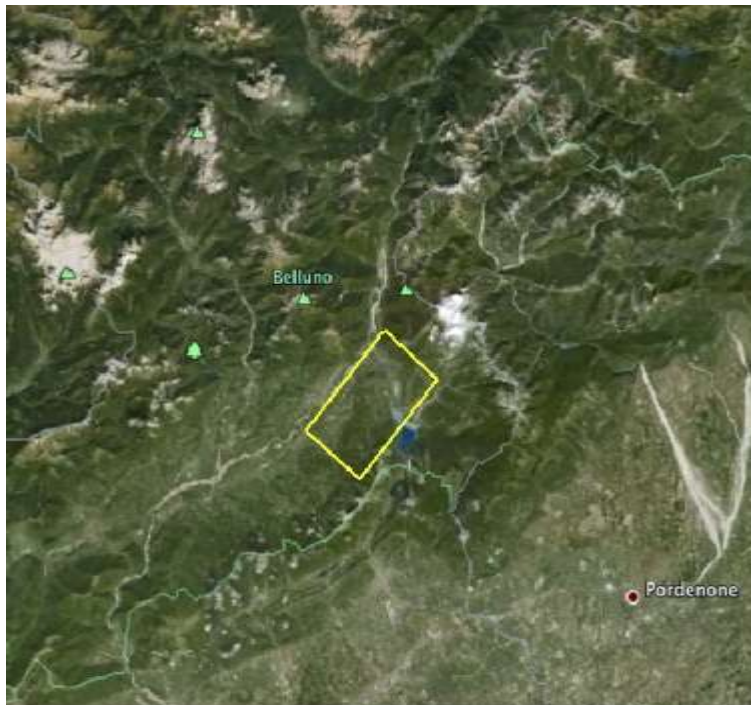


Rappresentazione artistica del terremoto di Basel 1356



Intensità

Terremoto di Belluno del 29 giugno 1873, Mw 6.3

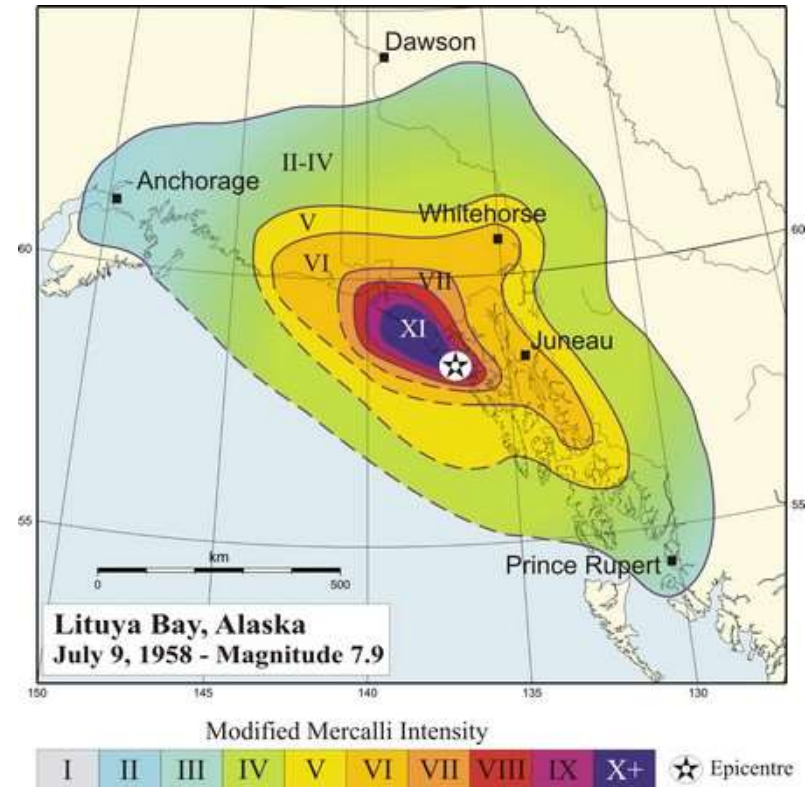
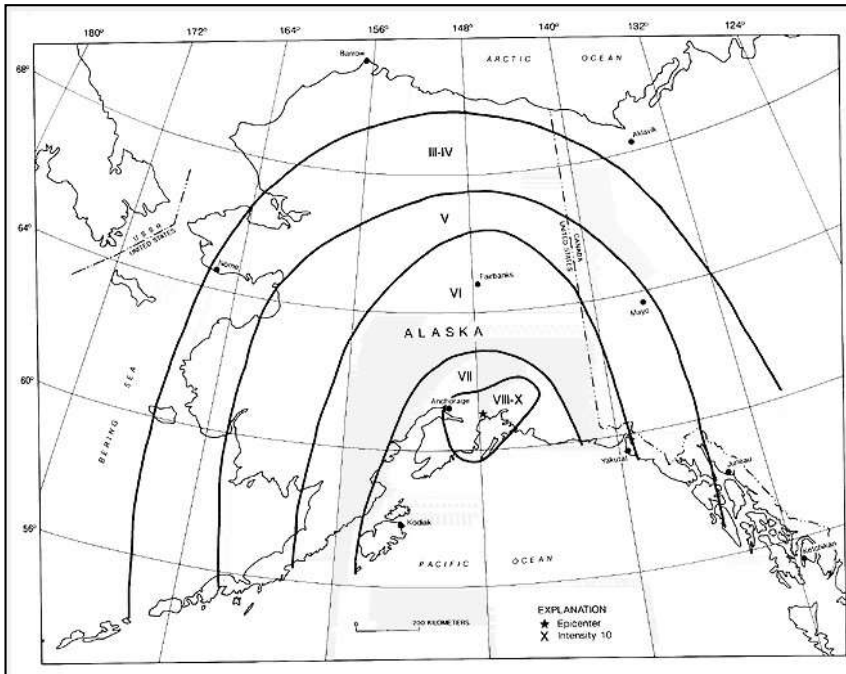




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- In order to evaluate the active seismicity of a region questionnaires may be distributed to the population, asking for observations that can be used to estimate the intensity experienced. The questionnaires are evaluated with the aid of an intensity scale, and the intensity recorded at the location of each observer is plotted on a map.
- Continuous lines are then drawn to outline places with the same intensity, in the same way that contour lines are used on topographic maps to show elevation.





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- Comparison of the isoseismal maps with geological maps helps explain the response of the ground to the shake of an earthquake. This is valuable information for understanding earthquake risk. The foundation on which structures are erected plays a vital role in their survival of an earthquake. For example, soft sediments can amplify the ground motion, enhancing the damage caused. This is even more serious when the sediments have a high water content, in which case liquefaction of the sediments can occur, robbing structures built on them of support and promoting their collapse.

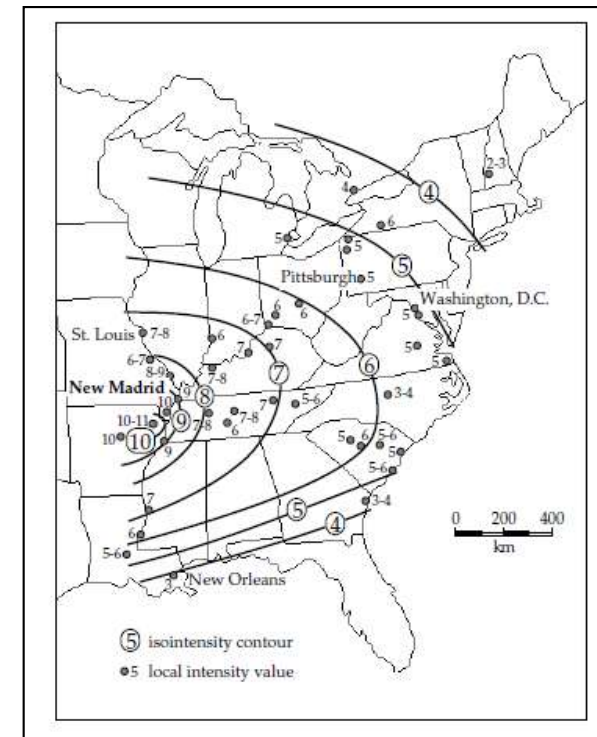


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- Intensity data play an important role in determining the historic seismicity of a region. An earthquake has dramatic consequences for a population; this was especially the case in the historic past, when real hazards were augmented by superstition.
- The date (and even the time) of occurrence of strong earthquakes and observations of their local effects have been recorded for centuries in church and civil documents. From such records it is sometimes possible to extract enough information for a given earthquake to estimate the intensity experienced by the observer.
- If the population density is high enough, it may be possible to construct an isoseismal map from which the epicenter of the tremor may be roughly located.

An interesting example of this kind of analysis is the study of the New Madrid earthquakes of 1811–1812, which caused devastation in the Mississippi valley and were felt as far away as the coastlines along the Atlantic and Gulf of Mexico. There were probably three large earthquakes, but the events occurred before the invention of the seismograph so details of what happened are dependent on the subjective reports of observers. Historical records of the era allow development of an intensity map for the settled area east of the Mississippi, but the pioneering population west of the river was at that time too sparse to leave adequate records for intensity interpretation. On the basis of the available evidence these earthquakes are estimated to have had magnitudes of 7.8–8.1.





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Danni alla chiesetta di Buia dovuti al terremoto del Friuli 1976



Effetti di sito





EFFETTI LOCALI INDOTTI DA UN EVENTO SISMICO

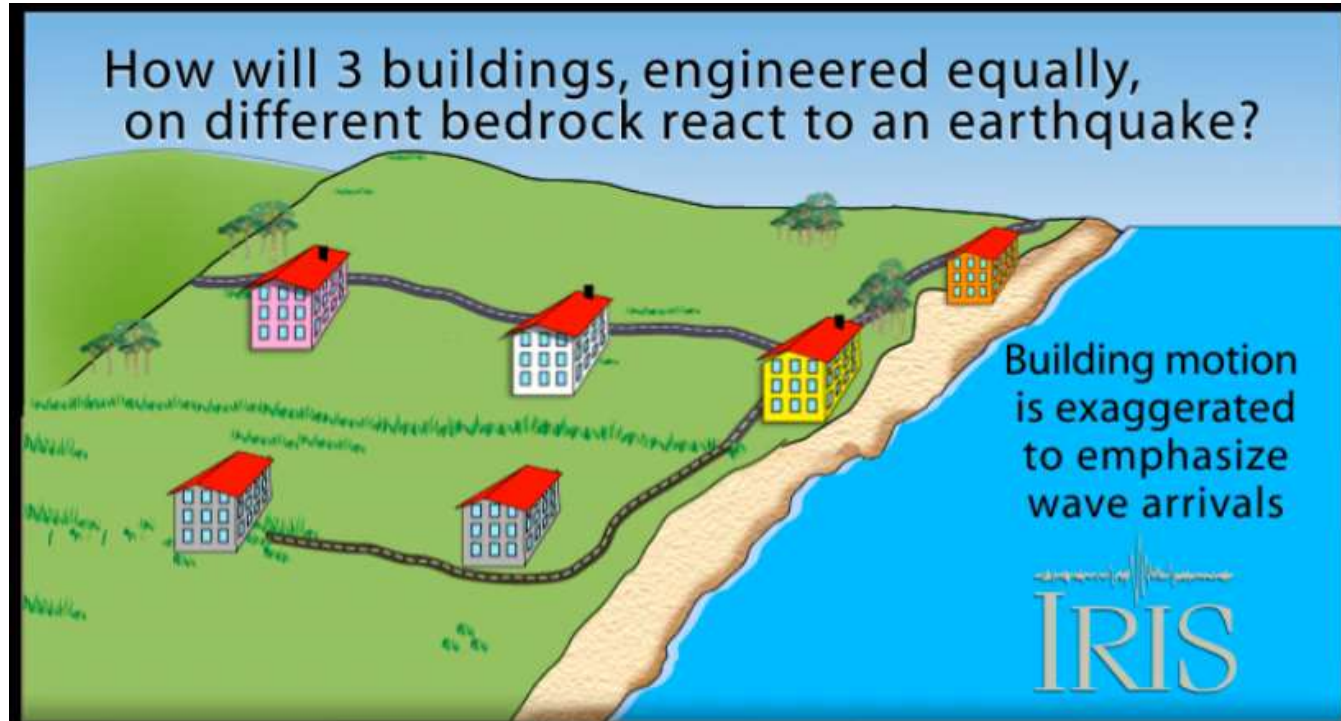


Emilia

20 maggio 2012 ore 04:03

M=5.9

Effetti di sito

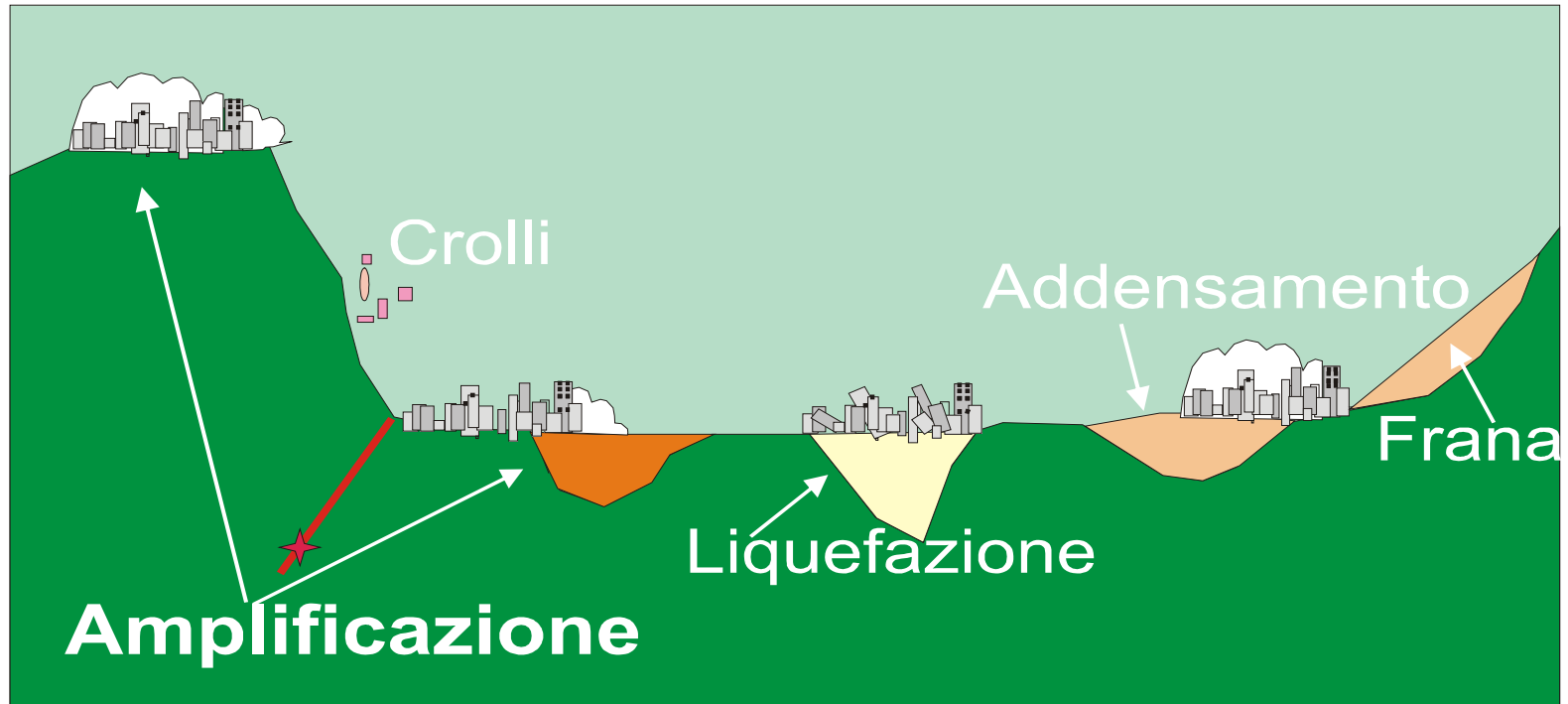


Two variables affect damage during earthquake:

- 1) Intensity of shaking (*felt motion, not magnitude*)
- 2) Engineering



EFFETTI LOCALI INDOTTI DA UN EVENTO SISMICO



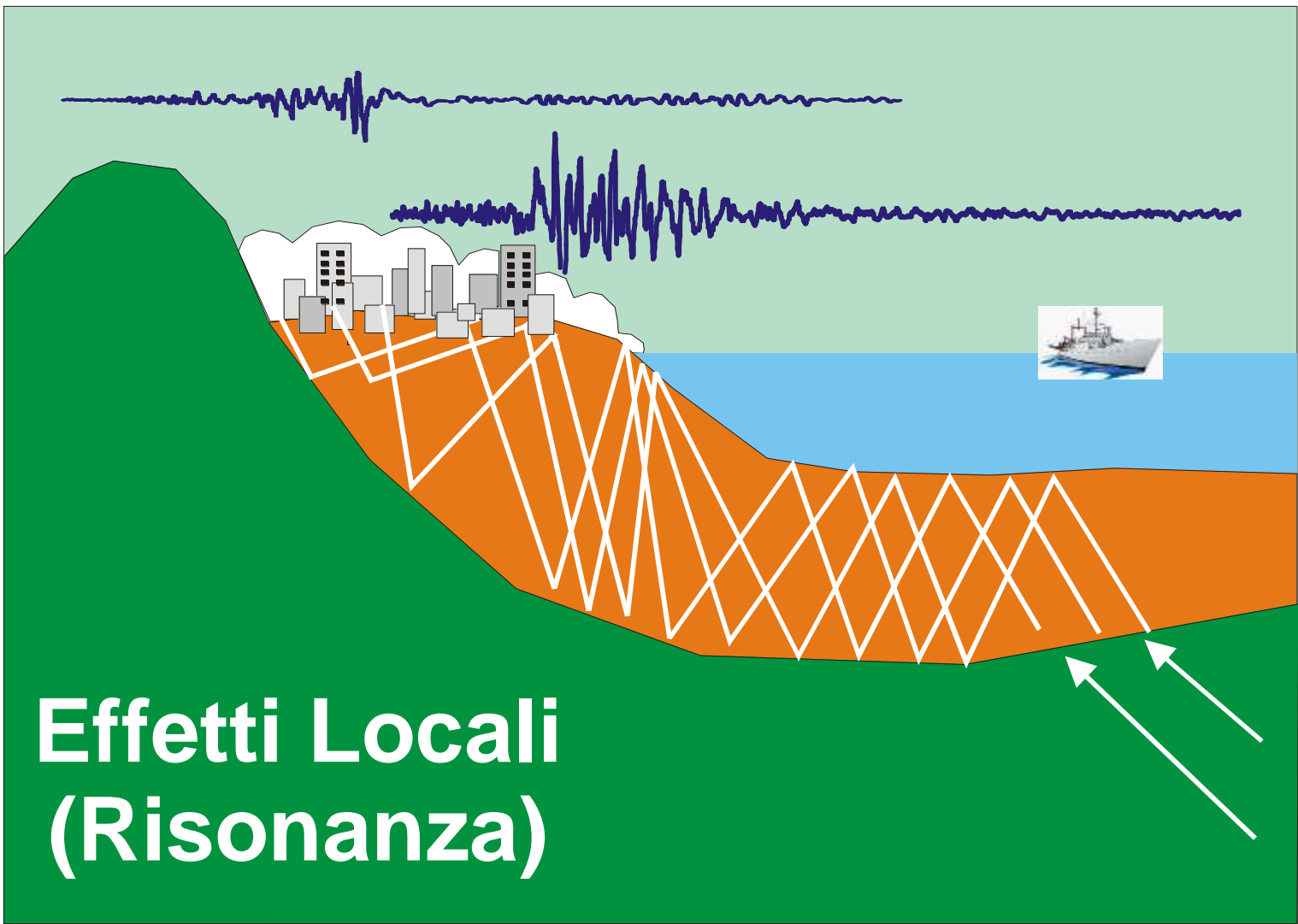


EFFETTI

Fisica Terrestre 2022-2023

LOCALI INDOTTI DA UN EVENTO SISMICO

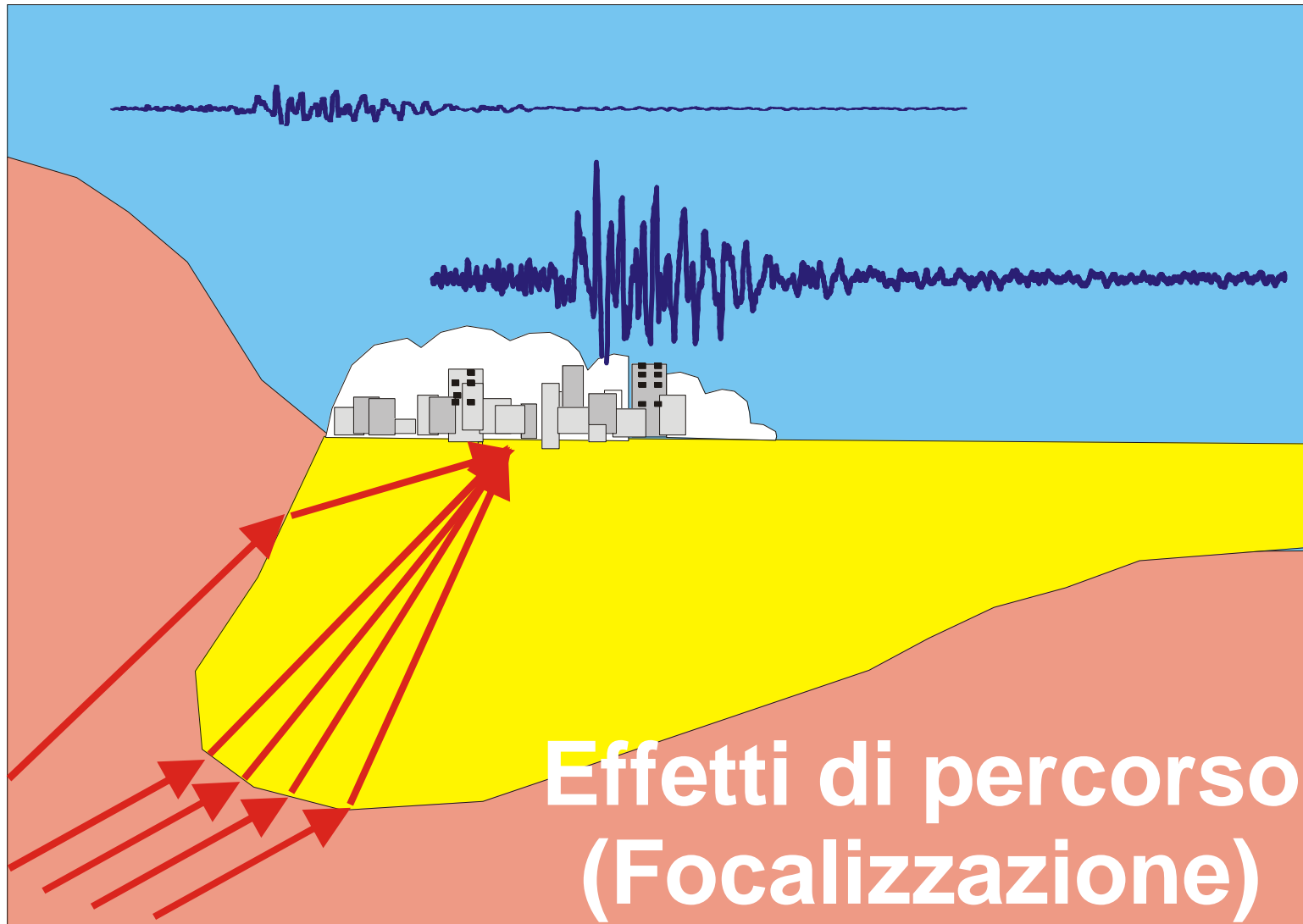
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**Effetti Locali
(Risonanza)**

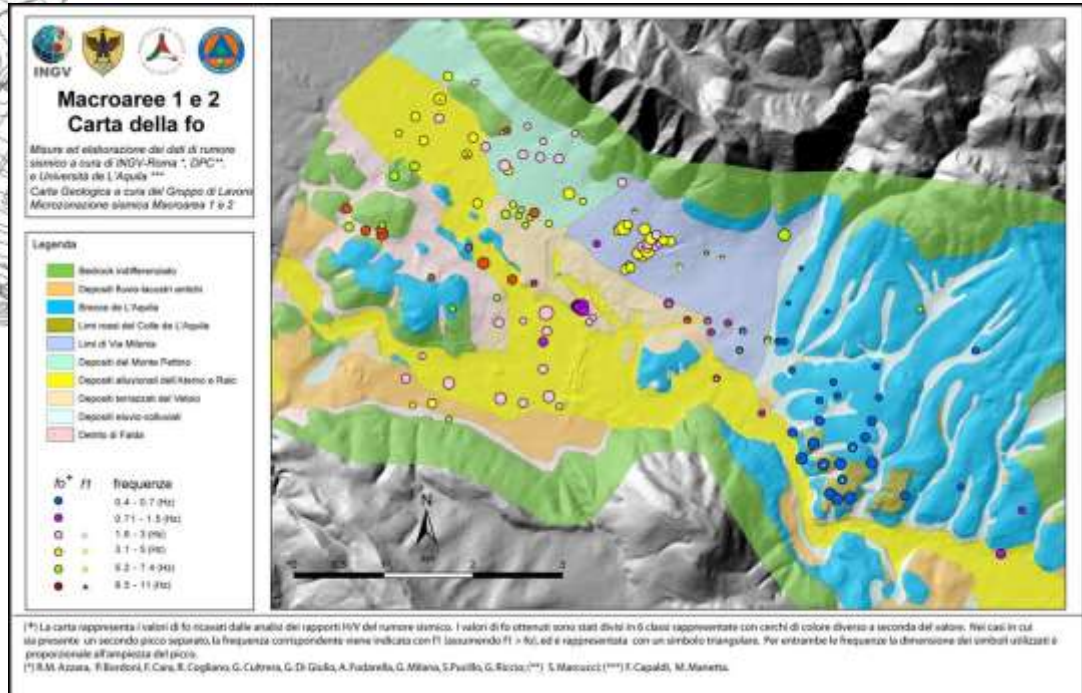
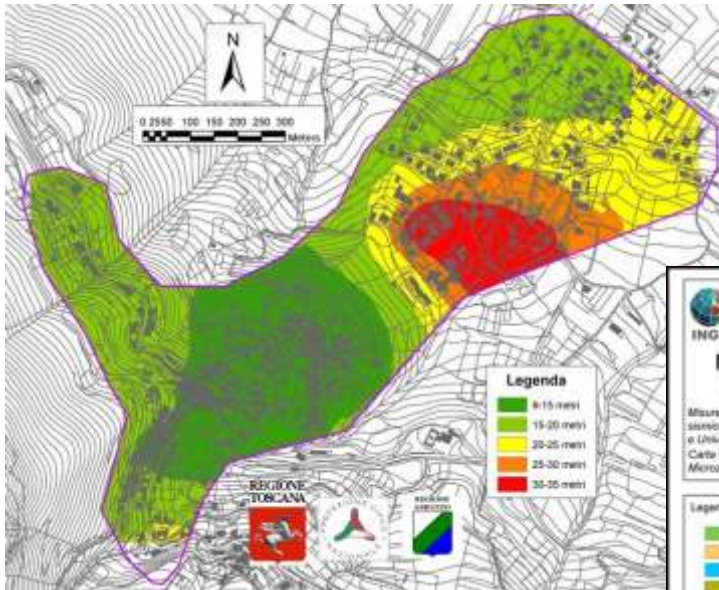


EFFETTI LOCALI INDOTTI DA UN EVENTO SISMICO





Microzonazione sismica



[*] La carta rappresenta i valori di fo ricavati dalle analisi dei rapporti HVV del rumore sismico. I valori di fo ottenuti sono stati divisi in 6 classi rappresentate con cerchi di colore diverso a seconda del valore. Nei casi in cui sia presente un secondo picco separato, la frequenza corrispondente viene indicata con f1. Assumendo f1 > fo, ed è rappresentata con un simbolo triangolare. Per entrambe le frequenze la dimensione dei simboli utilizzati è proporzionale all'ampiezza del picco.

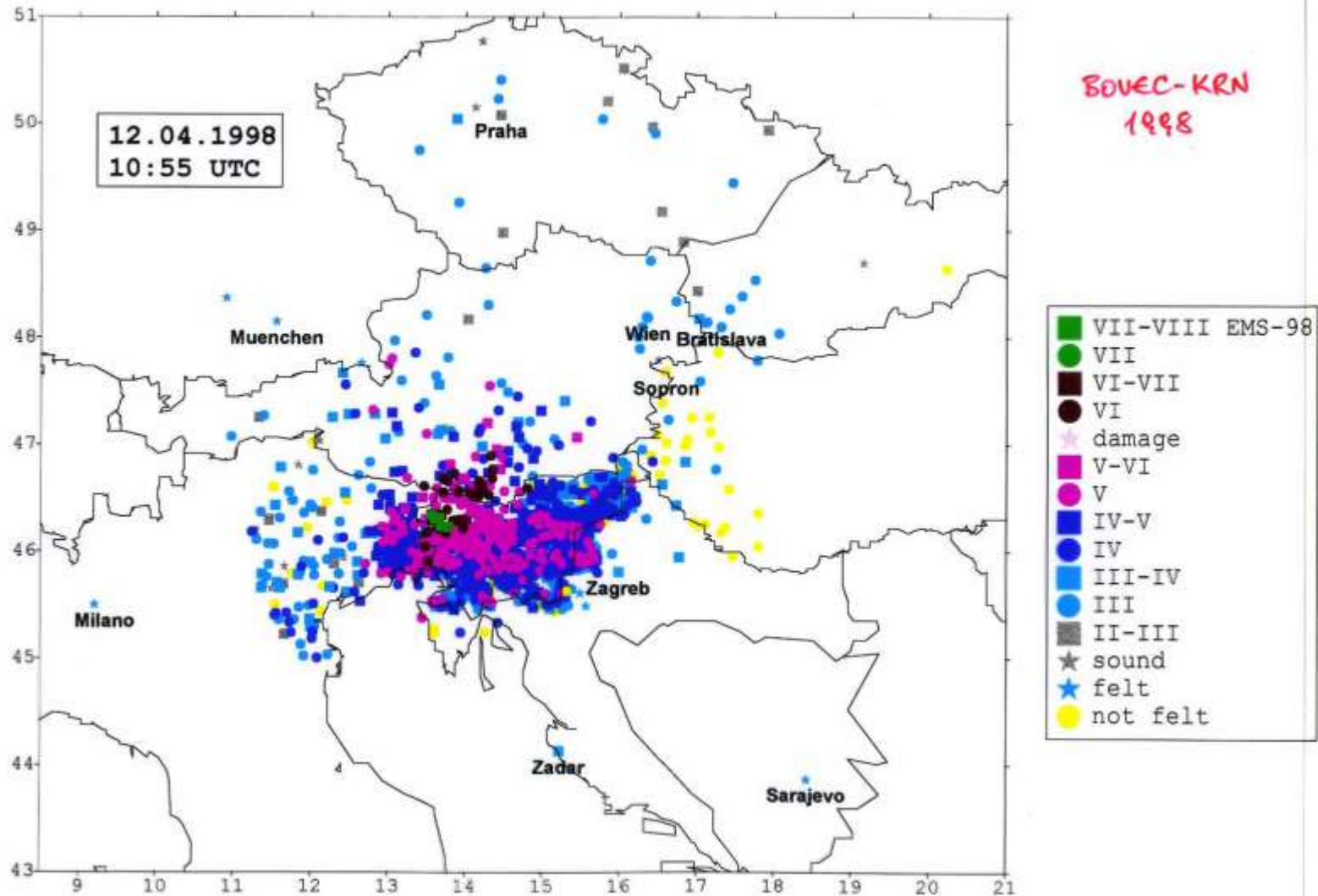
[**] I.R.M. Azara, F. Birelli, F. Carr, R. Cogliaro, G. Cutrona, G. Di Giulio, A. Fustarelli, G. Milano, S. Pavullo, G. Ricci, S. Marucci
 [***] I.R.M. Azara, F. Birelli, F. Carr, R. Cogliaro, G. Cutrona, G. Di Giulio, A. Fustarelli, G. Milano, S. Pavullo, G. Ricci, S. Marucci



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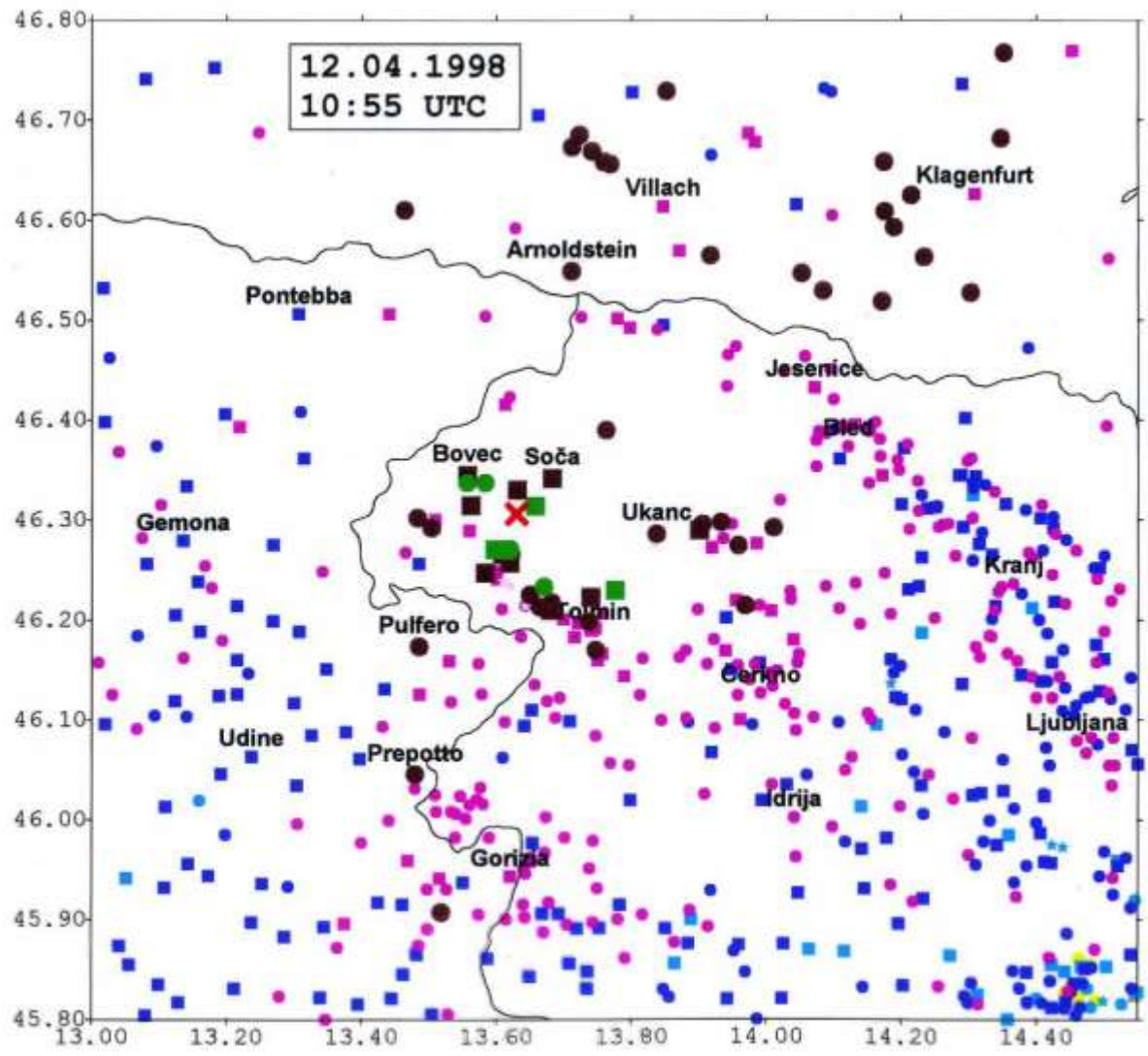
INTENSITIES OF THE MAIN SHOCK





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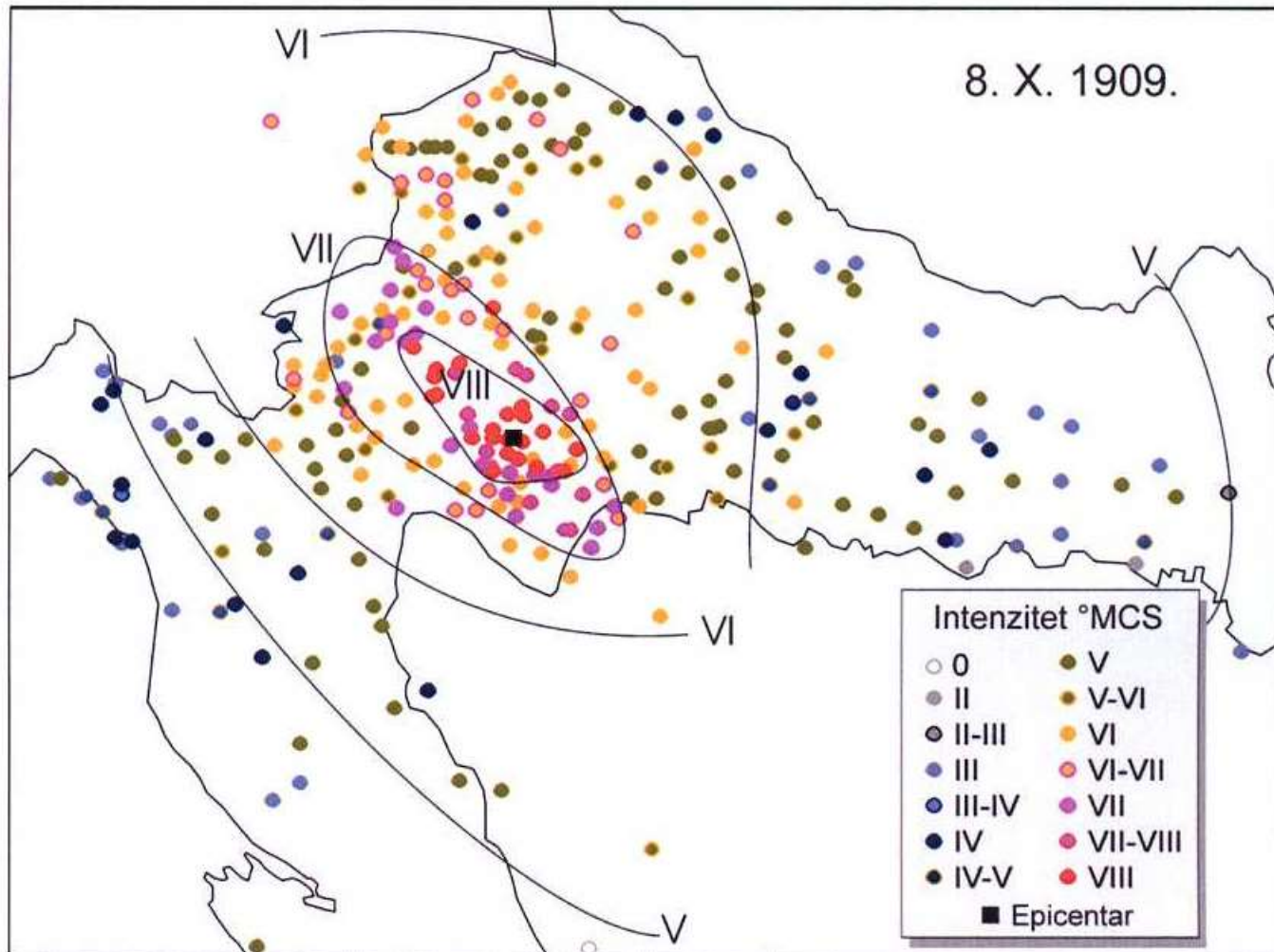
TERREMOTO DI
BOVEC-KRNJ
(SLOVENIA)
1998

- ✗ epicentre
- VII-VIII EMS-98
- VII
- VI-VII
- VI
- ★ damage
- V-VI
- V
- IV-V
- IV
- III-IV
- III
- II-III
- ★ sound
- ★ felt
- not felt



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Tracciamento delle isosisme



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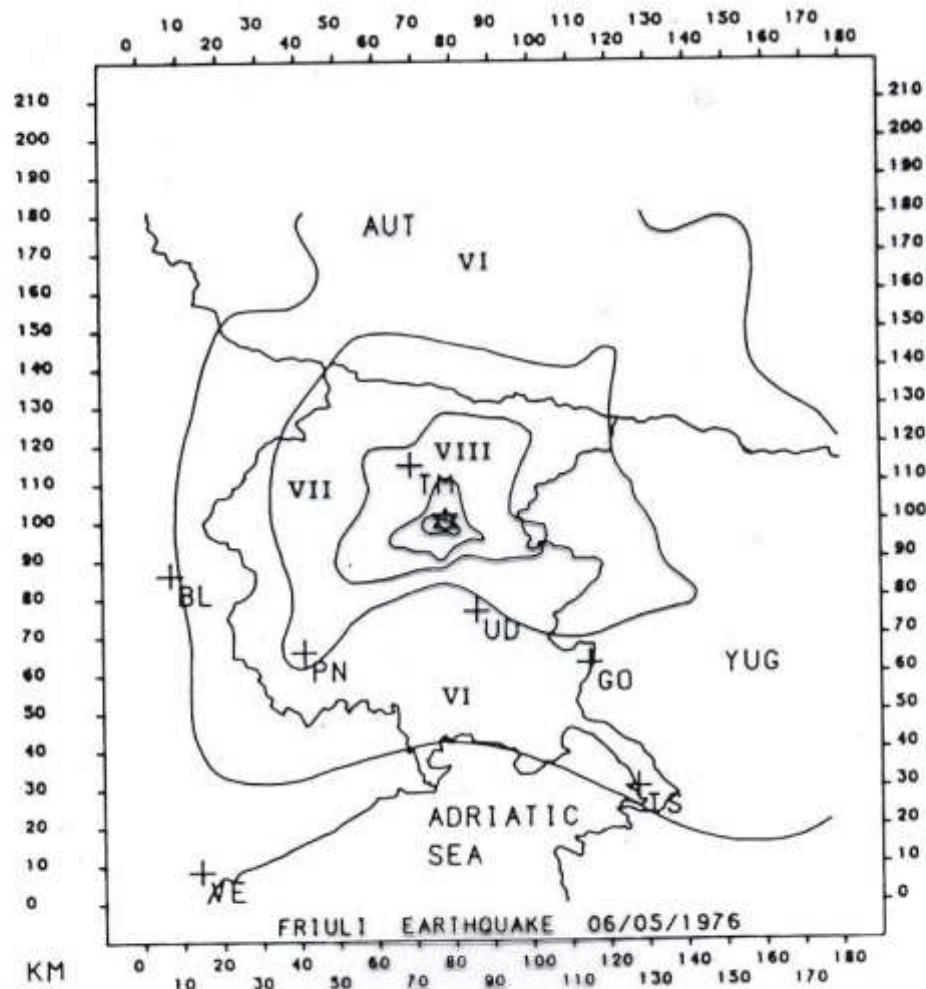


Figure 10. Observed isoseismals of the May 6, 1976 Friuli earthquake, MSK scale (after Giorgetti, 1976).



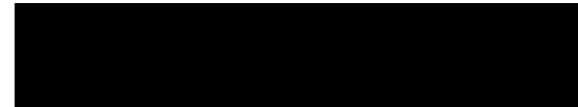
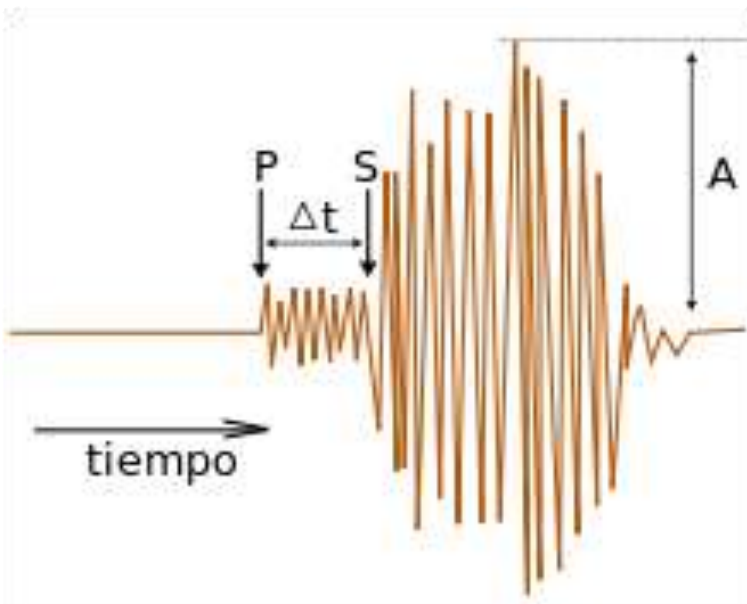
Earthquake: magnitude

- Magnitude is an experimentally determined measure of the “size” of an earthquake. In 1935 C. F. Richter attempted to grade the sizes of local earthquakes in Southern California on the basis of the amplitude of the ground vibrations they produced at a known distance from the epicenter. The vibrations were recorded by seismographs, which were standardized to have the same response to a given stimulus.
- Richter’s original definition of magnitude was based on surface-wave amplitudes (A_s) recorded by seismographs at an epicentral distance of 100 km.
- Because seismographs were located at various distance from the earthquake, an extra term was added to compensate for attenuation of the signal with increasing epicentral distance.



Magnitudo

La **magnitudo**, più propriamente, è un parametro atto a rappresentare l'energia sprigionata da un terremoto. Viene calcolata sulla base di misure effettuate sul sismogramma, maggiore è l'energia liberata dal sisma maggiori sono le ampiezze delle oscillazioni registrate. Misurata per molti anni (simbolo M) tramite la Scala Richter, oggi esistono diverse scale di magnitudo. Tutte le scale, a differenza della Scala Mercalli, non sono decimali ma *logaritmiche* perciò la differenza di un'unità implica un potenziale energetico di circa *30 volte maggiore o minore*.

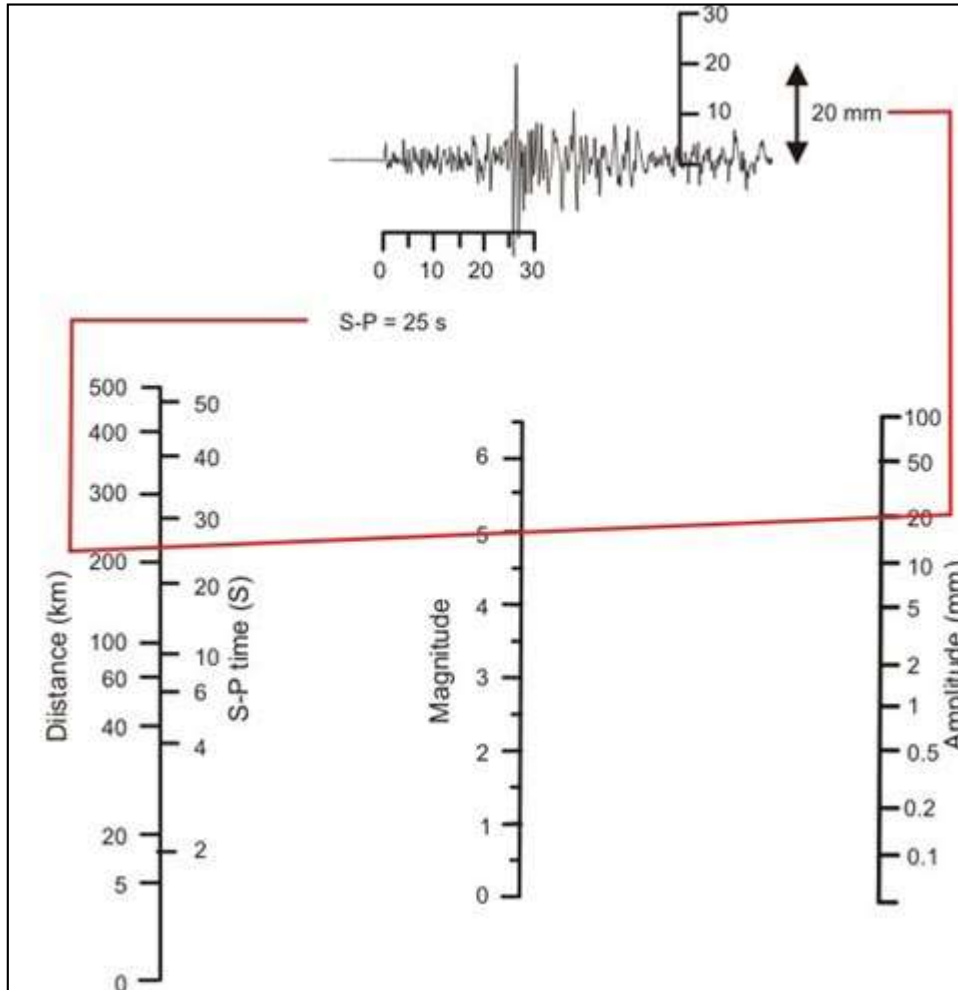


A = ampiezza massima delle oscillazioni del terremoto che si sta osservando
A₀ = ampiezza massima delle oscillazioni causate da un terremoto di riferimento
ε = fattore di correzione (distanza reale dall'epicentro, profondità dell'ipocentro)



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Procedure for calculating the local magnitude, M_L

1. Measure the distance to the focus using the time interval between the S and the P waves ($S - P = 24$ seconds).
2. Measure the height of the maximum wave motion on the seismogram (23 millimeters).
3. Place a straight edge between appropriate points on the distance (left) and amplitude (right) scales to obtain magnitude $M_L = 5.0$.

$$M_L = \log_{10}(A) - \log_{10}(\Delta) \quad [Eq 3]$$

General form of Magnitude scales:

$$M = \log(A/T) + F(h, \Delta) + C$$

A is the amplitude of the signal

T is its dominant period

F is a correction for the variation of amplitude with the earthquake's depth h and distance Δ from the seismometer

C is a regional scale factor



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- Increasingly sensitive instruments allowed the recording of signals from distant earthquakes; those from events with epicentral distances greater than 20° are known as teleseismic signals. Originally, the magnitude was determined from the horizontal ground motion, because seismological stations were equipped mainly with horizontal-motion seismometers. However, the surface waves recorded by these instruments consist of superposed Love and Rayleigh waves, which complicates the theoretical interpretation of the records.
- Vertical-motion seismometers record only the Rayleigh waves (together with P- and SV-waves), and so progressively the definition of surface-wave magnitude has come to be based on the vertical component of motion. The majority of surface-wave magnitudes assigned to earthquakes worldwide are now based on vertical-motion records.
- The International Association for Seismology and Physics of the Earth's Interior (IASPEI) has adopted the following definition of the surface-wave magnitude (M_s) of an earthquake:

$$M_s = \log_{10} \left(\frac{A_s}{T} \right) + 1.66 \log_{10} (\Delta) + 3.3 \quad \text{119 [Eq 4]}$$

$$M_s = \log_{10} \left(\frac{A_s}{T} \right) + 1.66 \log_{10} (\Delta) + 3.3 \quad [Eq 4]$$

where A_s is the vertical component of the ground motion in micrometers (μm) determined from the maximum Rayleigh-wave amplitude, T is the period of the wave (18–22 seconds), Δ is the epicentral distance in degrees ($20^\circ < \Delta < 160^\circ$), and where the earthquake has a focal depth of less than 50 km.

- The depth of the source affects the nature of the seismic wave train, even when the same energy is released. An earthquake with a deep focus may generate only a small surface-wave train, while shallow earthquakes cause very strong surface waves. Equation 4 for M_s was derived from the study of shallow earthquakes, observed at a distance greater than 20° . Therefore, corrections must be made to the computed value of M_s to compensate for the effects of a focal depth greater than 50 km or epicentral distance less than 20°

The amplitude of body waves is not sensitive to the focal depth. As a result, earthquake magnitude scales have also been developed for use with body waves. An equation, proposed by B. Gutenberg in 1945, can be used to calculate a body-wave magnitude (m_b) from the maximum amplitude (A_p) of the ground motion associated with P-waves having a period (T) of less than 3 s:

$$m_b = \log_{10} \left(\frac{A_p}{T} \right) + Q(\Delta, h) \quad [Eq 5]_{20}$$



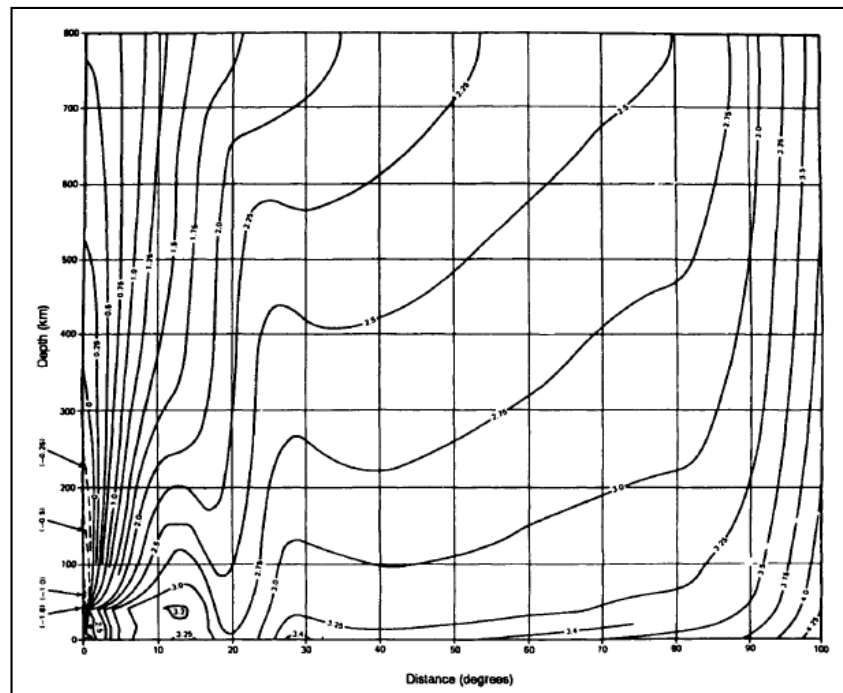
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$$M_S = \log_{10} \left(\frac{A_s}{T} \right) + 1.66 \log_{10} (\Delta) + 3.3 \quad [Eq 4]$$

$$m_b = \log_{10} \left(\frac{A_p}{T} \right) + Q(\Delta, h) \quad [Eq 5]$$

Where $Q(\Delta, h)$ is an empirical correction for signal attenuation due to epicentral distance (Δ) and focal depth (h) that is made by reading directly from a graph or table of values.

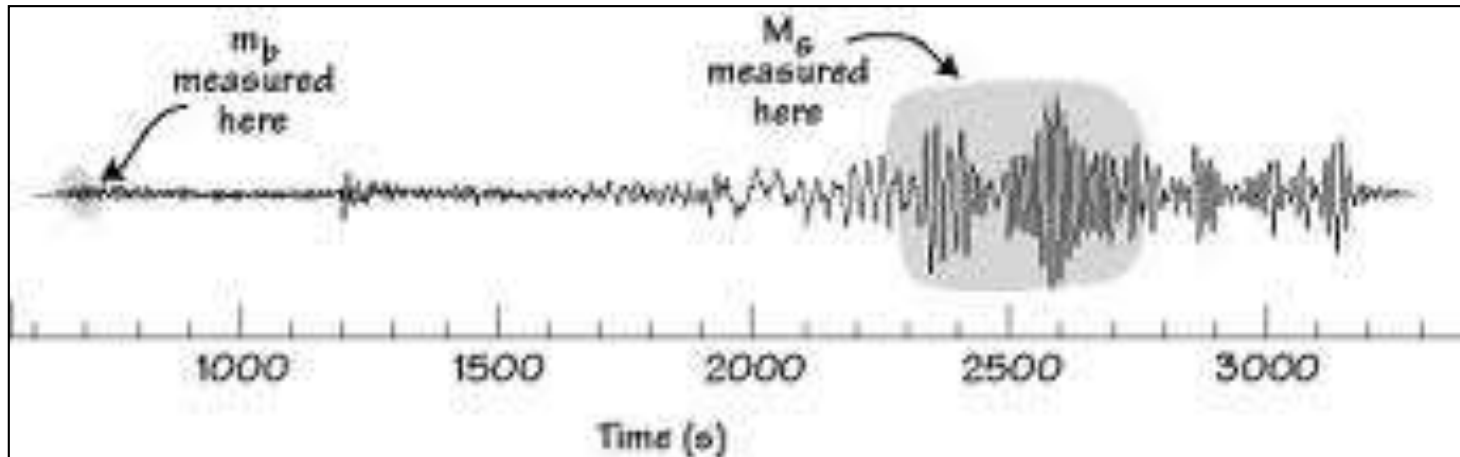




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For some earthquakes both M_s and m_b can be calculated. Unfortunately, the different estimates of magnitude often do not agree well, except for small earthquakes. This is due to the way the ground responds to a seismic event, and to the different natures of body waves and surface waves. Body waves have a different dependence of amplitude on frequency than do surface waves. m_b is estimated from a highfrequency (1 Hz) phase whereas M_s is determined from low-frequency (0.05 Hz) vibrations.





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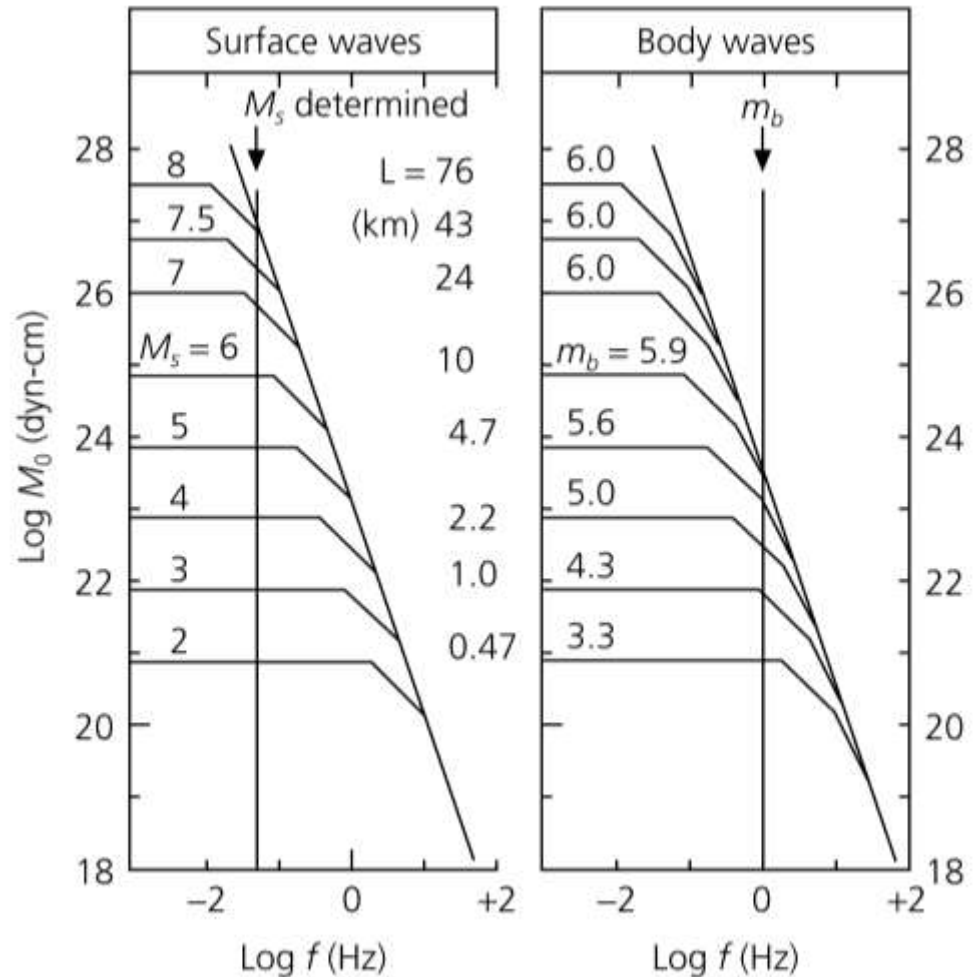
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DIFFERENT MAGNITUDES REFLECT ENERGY RELEASE AT DIFFERENT PERIODS

1 s - Body wave
magnitude m_b

20 s - Surface wave
magnitude M_s

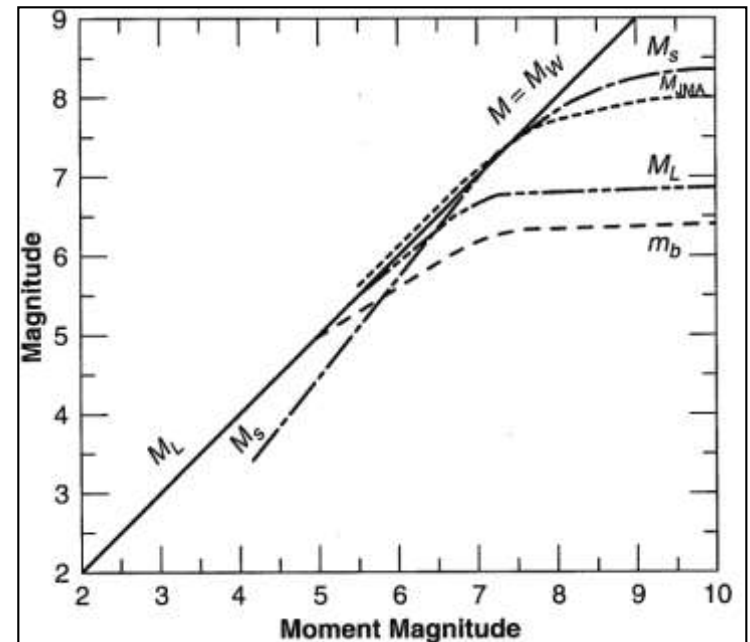
Long period - moment
magnitude M_w derived
from moment M_0





Earthquake: magnitude saturation

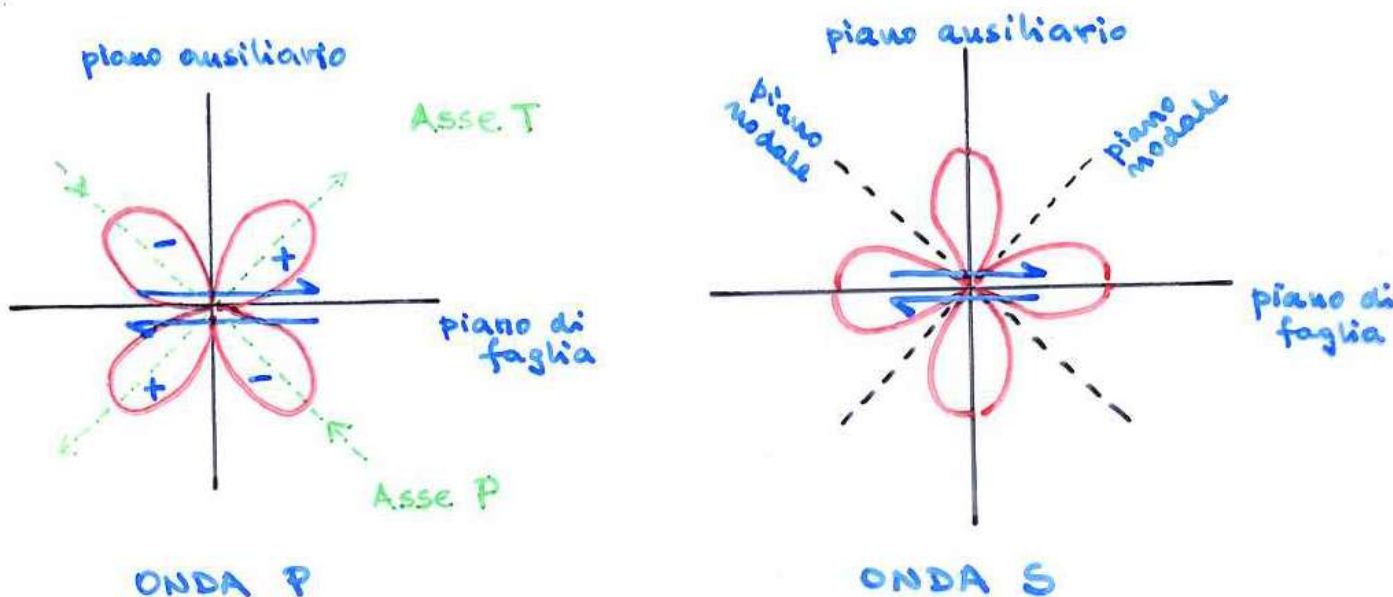
- Above a certain size, each method becomes insensitive to the size of the earthquake, and exhibits magnitude saturation. This occurs for body-wave estimates at around $m_b > 6$; all larger earthquakes give the same body-wave magnitude. Similarly, surface-wave estimates of magnitude saturate at $M_s > 8$.
- Thus, for very large earthquakes, M_s and m_b underestimate the energy released. An alternative definition of magnitude, based upon the long-period spectrum (M_w) of the seismic wave, is preferred for very large earthquakes. It makes use of the physical dimensions of the focus.





PROFILI RADIAZIONE

Sia M_s che M_w tendono a saturare (non danno valori abbastanza alti) per terremoti molto grandi e dipendono dalla frequenza. Inoltre, I terremoti non irradiano onde sismiche uniformemente in tutte le direzioni, ma hanno profili di radiazione caratteristici:





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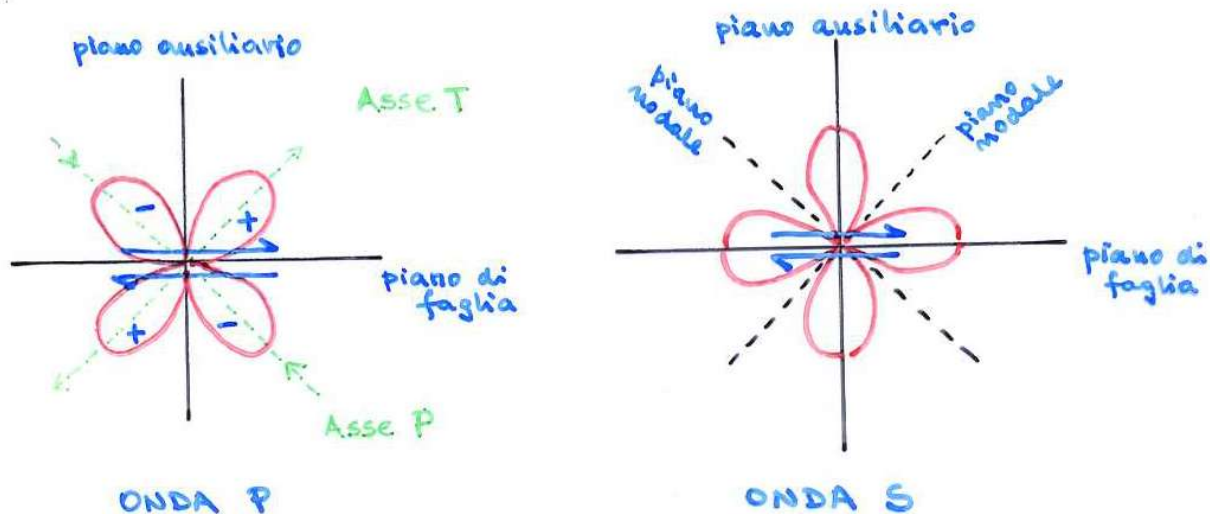
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Il terremoto non irradia onde P lungo la faglia e lungo la direzione perpendicolare ad essa (piani nodali), mentre per le onde S tali direzioni hanno il Massimo di radiazione. Per le onde P il Massimo di radiazione avviene nella direzione di massima compressione (con polarità primo impulso negativa) e di massima tensione (con polarità primo impulso positiva).

L'ampiezza dell'onda che arriva alla stazione dipende fortemente dall'azimuth della stazione rispetto alla faglia.

Una certa distribuzione di stazioni potrà quindi sovrastimare oppure sottostimare la magnitudine.

Per le ragioni sopra esposte si preferisce caratterizzare la entità del terremoto mediante un parametron più fisico, il momento sismico e con una magnitudine legata ad esso.





- As discussed in the elastic rebound model, a tectonic earthquake arises from abrupt displacement of a segment of a fault. The area S of the fractured segment and the amount by which it slipped D can be inferred.
- Together with the rigidity modulus m of the rocks adjacent to the fault, these quantities define the seismic moment M_0 of the earthquake. Assuming that the displacement and rigidity are constant over the area of the rupture:

$$M_0 = \mu SD \quad [Eq 6]$$

- The seismic moment can be used to define a **moment magnitude** (M_w). The definition adopted by the responsible commission of IASPEI is:

$$M_w = \frac{2}{3} (\log_{10} M_0 - 9.1) \quad [Eq 7]$$

In this equation M_0 is in N m

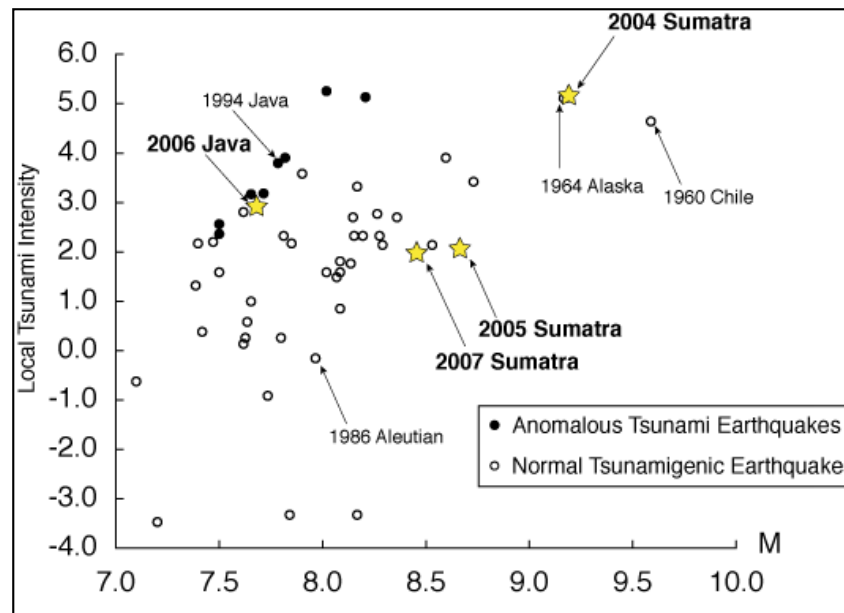
- M_w is more appropriate for describing the magnitudes of very large earthquakes. It has largely replaced M_s in scientific evaluation of earthquake size, although M_s is often quoted in reports in the media.



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- The magnitude scale is, in principle, open ended. Negative Richter magnitudes are possible, but the limit of sensitivity of seismographs is around -2. The maximum possible magnitude is limited by the shear strength of the crust and upper mantle, and since the beginning of instrumental recording none has been observed with a surface wave magnitude M_s as high as 9. However, this is largely due to a saturation effect resulting from the method by which surface-wave magnitudes M_s are computed.
- Seismic moment magnitudes M_w of 9 and larger have been computed for some giant earthquakes. The largest in recorded history was the 1960 Chile earthquake with $M_w=9.5$.



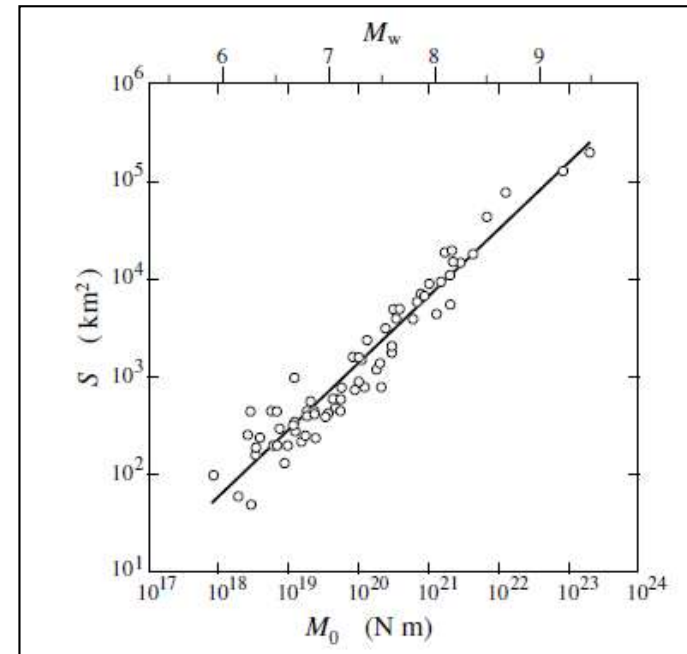


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- When an earthquake occurs on a fault, the ruptured area spreads in size from the point of initial failure, akin to opening a zip fastener. If the fault ruptures along a length L , and the fractured segment has a down-dip dimension w (referred to as the width of the faulted zone), the area S is equal to wL.

- This inference is supported by the correlation between seismic moment and the ruptured area S for many shallow earthquakes





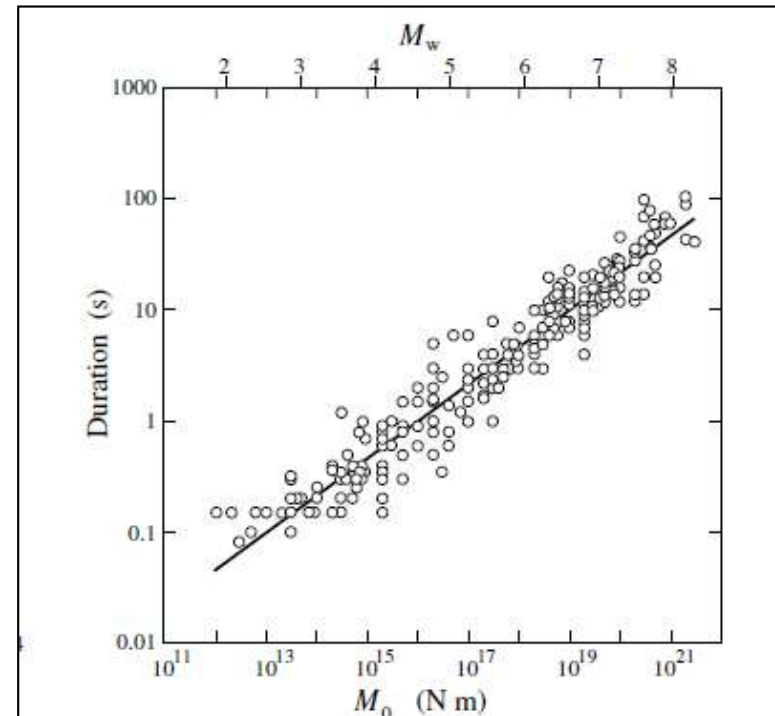
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- The length L of the rupture zone determines how long the ground motion lasts in an earthquake. This factor has a significant bearing on the extent of damage to structures during moderate to strong earthquakes. Theory gives the speed of propagation of the rupture as 75–95% of the shear wave velocity.
- Assuming that the rupture propagates along the fault at about $2.5\text{--}3.0 \text{ km s}^{-1}$, the duration of the ground motion can be estimated for an earthquake of a given magnitude or seismic moment

- For example, close to the epicenter of a magnitude 5 earthquake the vibration may last a few seconds, whereas in a magnitude 8 earthquake it might last about 50 s. However, there is a lot of scatter about the ideal correlation.

- In the great Sumatra earthquake, with magnitude 9.0, the rupture continued for some 500 s!





Earthquake: energy

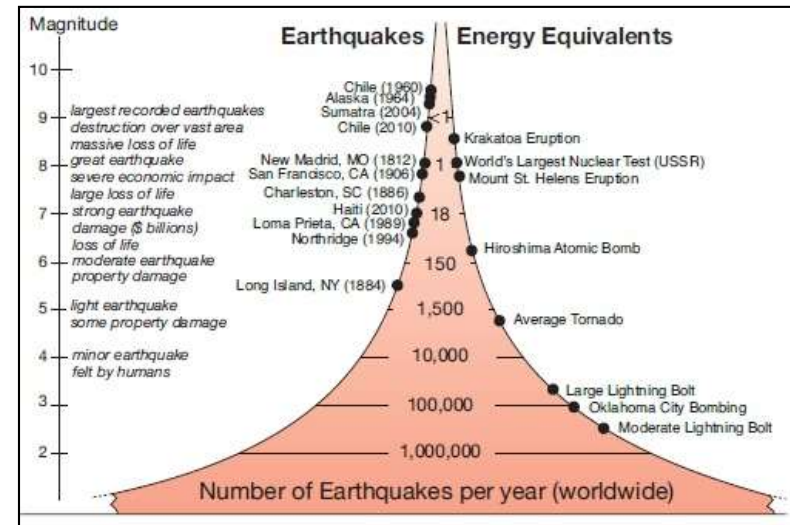
- The definition of earthquake magnitude relates it to the logarithm of the amplitude of a seismic disturbance. Noting that the energy of a wave is proportional to the square of its amplitude it should be no surprise that the magnitude is also related to the logarithm of the energy. Several equations have been proposed for this relationship.
- An empirical formula worked out by Gutenberg relates the energy release E to the surface-wave magnitude M_s :

$$\log_{10} E = 4.8 + 1.5M_s \quad [Eq\ 8]$$

• The logarithmic nature of the formula means that the energy release increases very rapidly with increasing magnitude. For example, when the magnitudes of two earthquakes differ by 1, their corresponding energies differ by a factor 32 ($=10^{1.5}$). Hence, a magnitude 7 earthquake releases about 1000 times the energy of a magnitude 5 earthquake. Another way of regarding this observation is that it takes 1000 magnitude 5 earthquakes to release the same amount of energy as a single large earthquake with magnitude 7.

Earthquake magnitude	Number per year	Annual energy [10^{15} J yr ⁻¹]
≥ 8.0	≈ 1	≈ 100
7-7.9	17	190
6-6.9	134	45
5-5.9	1,319	14
4-4.9	$\approx 13,000$	4
3-3.9	$\approx 130,000$	1
2-2.9	$\approx 1,300,000$	0.4

- Multiplying the mean number of earthquakes per year by their estimated energy (using one of the energy–magnitude equations) gives an impression of the importance of very large earthquakes.
- $M_s > 7$ are responsible for most of the annual seismic energy. In a year in which a very large earthquake ($M_s > 8$) occurs, most of the annual seismic energy is released in that single event.



- It is rather difficult to appreciate the amount of energy released in an earthquake from the numerical magnitude alone. A few examples help illustrate the amounts of energy involved. Earthquakes with $M_s=1$ are so weak that they can only be recorded instrumentally; they are referred to as micro earthquakes. The energy associated with one of these events is equivalent to the kinetic energy of a medium sized automobile weighing 1.5 tons which is travelling at 130 km h^{-1} (80 m.p.h.).
- The energy released by explosives provides another means of comparison, although the conversion of energy into heat, light and shock waves is proportionately different in the two phenomena. One ton of the explosive trinitrotoluene (TNT) releases about 4.2×10^9 joules of energy. Equation 8 shows that the 11 kiloton atomic bomb which destroyed Hiroshima released about the same amount of energy as an earthquake with magnitude 5.9. The energy released in a 1 megaton nuclear explosion is equivalent to an earthquake with magnitude 7.2; the 2004 Sumatra earthquake (magnitude 9.0) released an amount of energy equivalent to the detonation of 475 megaton bombs.



Earthquake: frequency

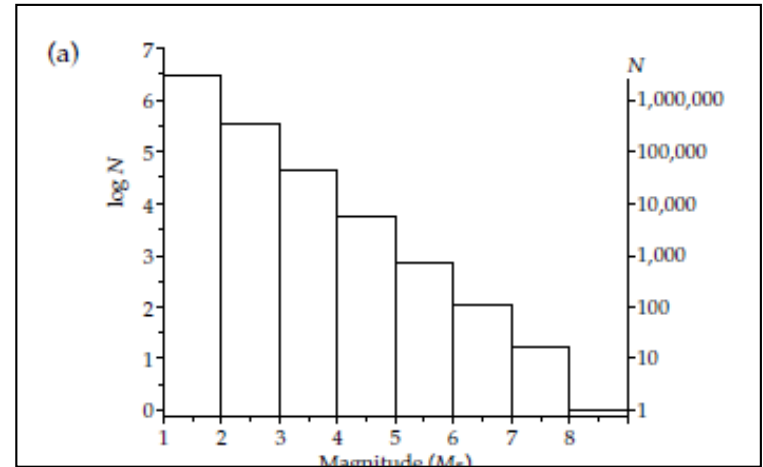
• Every year there are many small earthquakes, and only a few large ones. According to a compilation published by **Gutenberg and Richter in 1954**, the mean annual number of earthquakes in the years 1918–1945 with magnitudes 4–4.9 was around 6000, while there were only an average of about 100 earthquakes per year with magnitudes 6–6.9. The relationship between annual frequency (N) and magnitude (M_s) is logarithmic and is given by an equation of the form:

$$\log_{10} N = a - bM_s \quad [Eq 9]$$

- The value of a varies between about 8 and 9 from one region to another, while b is approximately unity for regional and global seismicity.
- Low values of b indicate area less fractured and characterize continental rift and deep earthquakes.
- High values of b indicate area highly fractured and are typical of oceanic ridges.
- The b value is determined with the method of maximum likelihood:

$$b = \frac{0.4343}{\overline{M} - M^*}$$

$$\begin{cases} \overline{M} = \text{magnitude average} \\ M^* = \text{smallest magnitude} - 0.05 \end{cases}$$



The Gutenberg Richter law characterizes a seismic area and is important for seismic hazard



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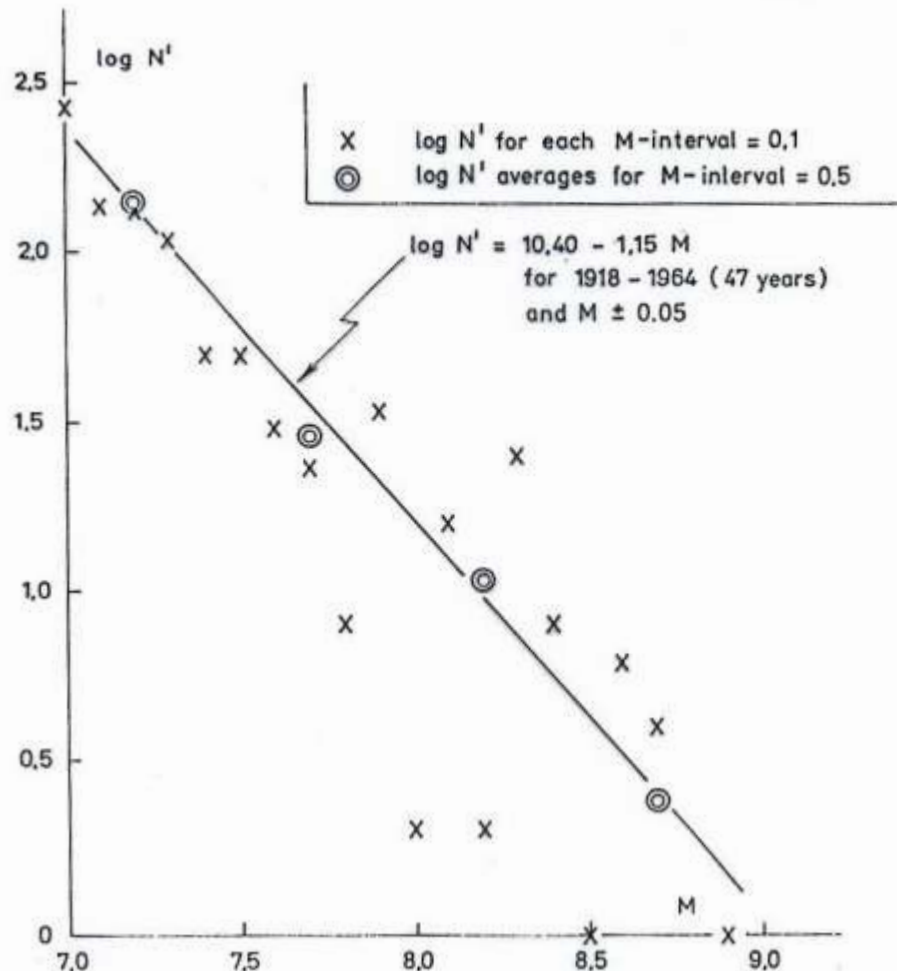
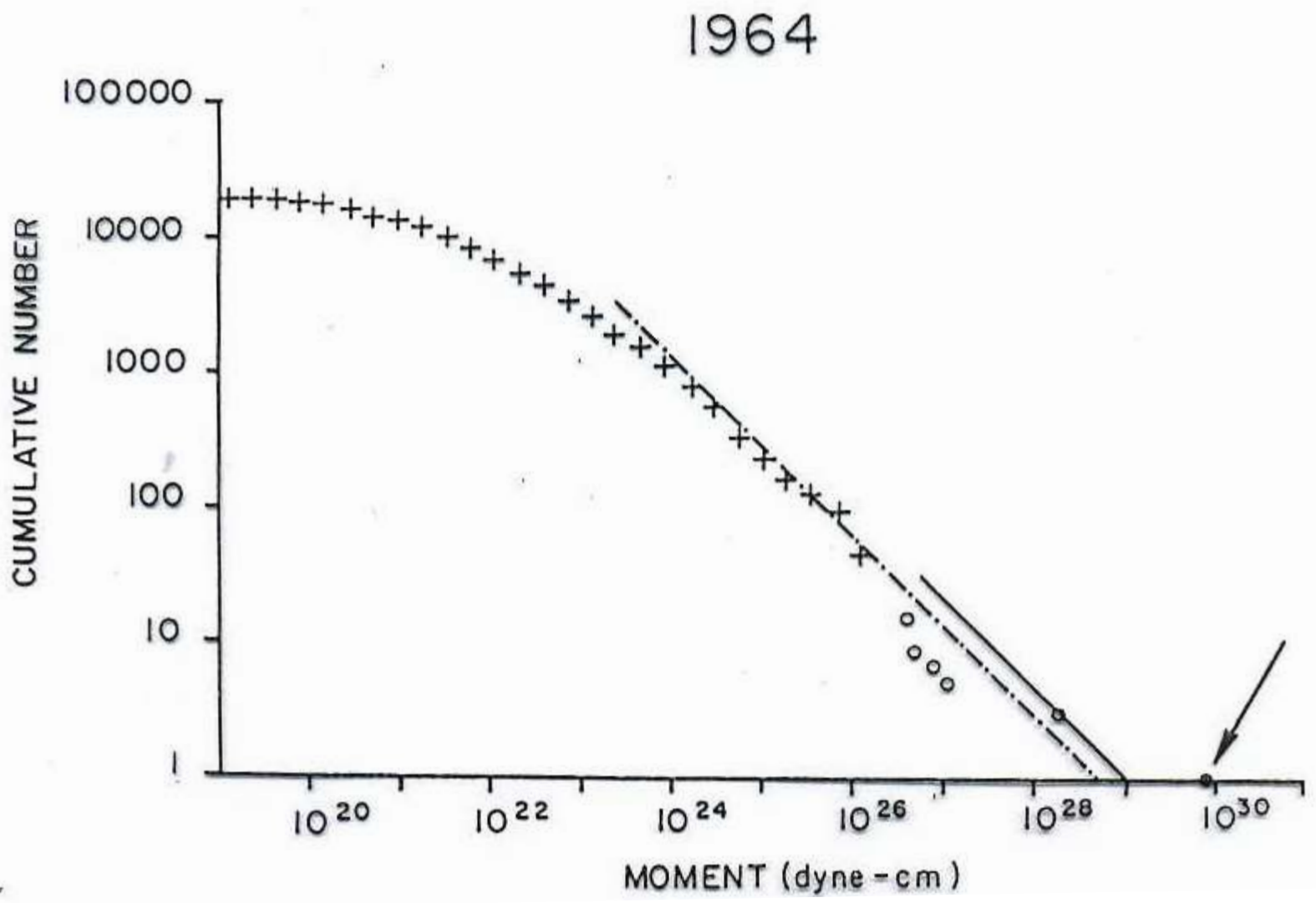


Fig. 49. Earthquake frequency for the earth as a whole for the interval 1918—64.



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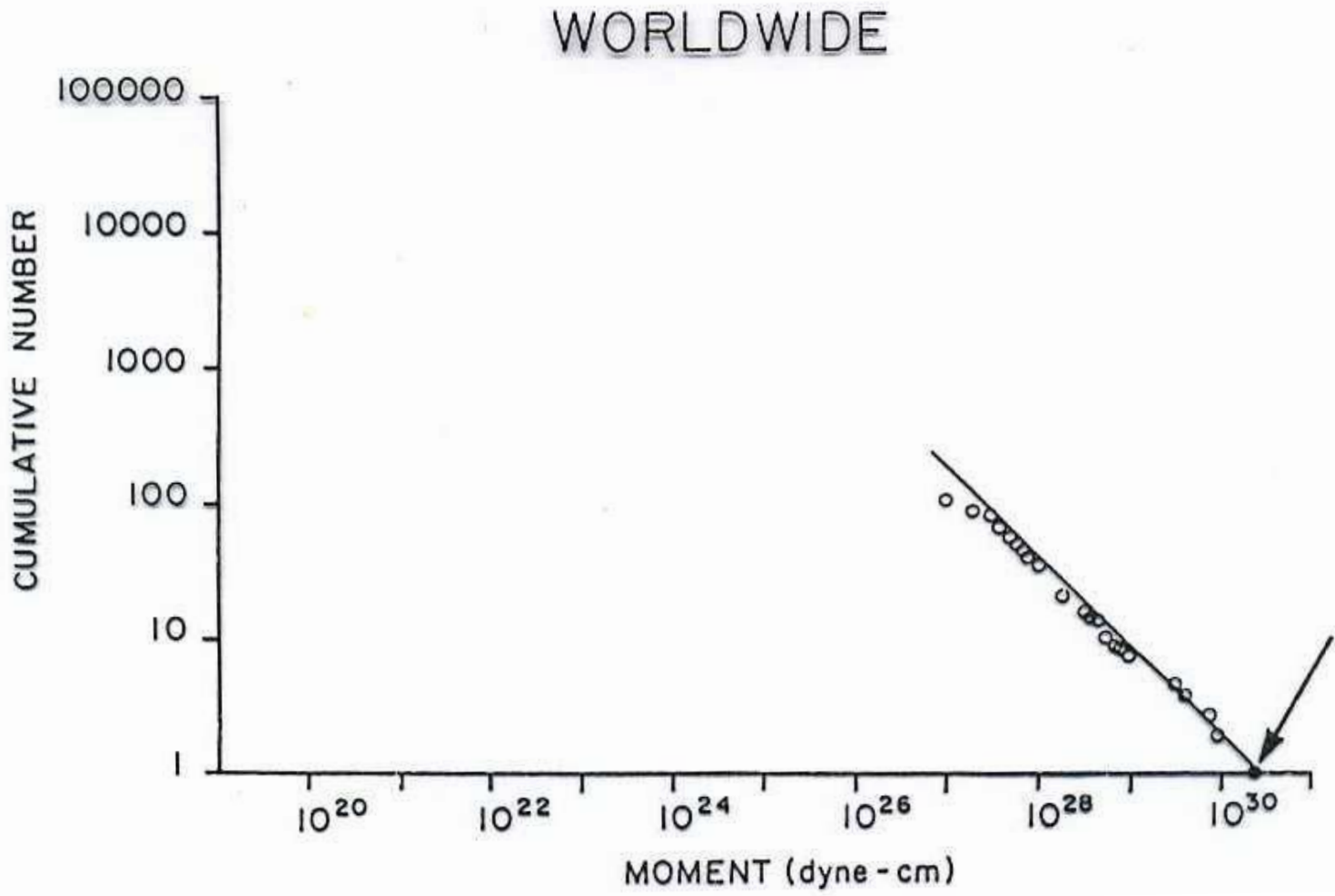
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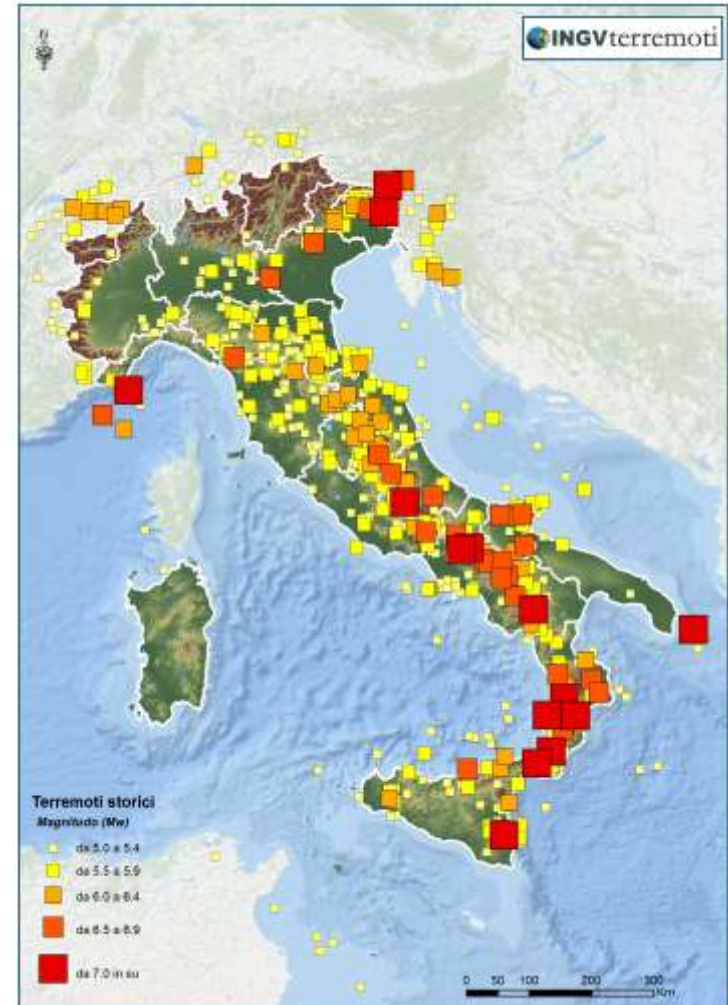
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Terremoti passati

Guardando la mappa dei terremoti italiani dall'anno 1000 al 2006, è facile notare che i **terremoti spesso avvengono in zone già colpite in passato**. Gli eventi storici più forti si sono verificati in Sicilia, nelle Alpi orientali e lungo gli Appennini centro-meridionali, dall'Abruzzo alla Calabria. Ma ci sono stati terremoti importanti anche nell'Appennino centro-settentrionale e nel Gargano. In particolare, **dal 1900 ad oggi si sono verificati 30 terremoti molto forti ($M_w \geq 5.8$)**, alcuni dei quali sono stati catastrofici. Qui di seguito li riportiamo in ordine cronologico. **Il più forte tra questi è il terremoto che nel 1908 distrusse Messina e Reggio Calabria.**

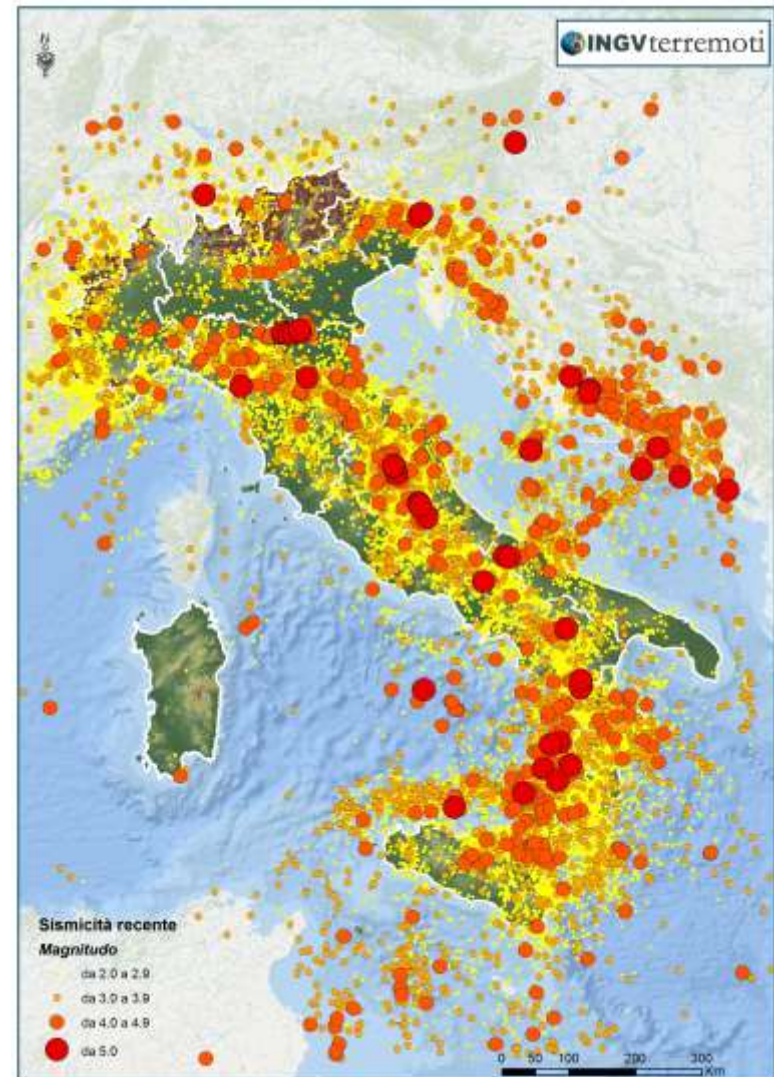
La sismicità dall'anno 1000 al 2006 by INGV



Terremoti recenti

Guardando la mappa degli ultimi 30 anni (1985-2014) di sismicità si nota che i terremoti recenti sono localizzati in aree distribuite principalmente lungo la fascia al di sotto degli Appennini, dell'arco Calabro e delle Alpi.

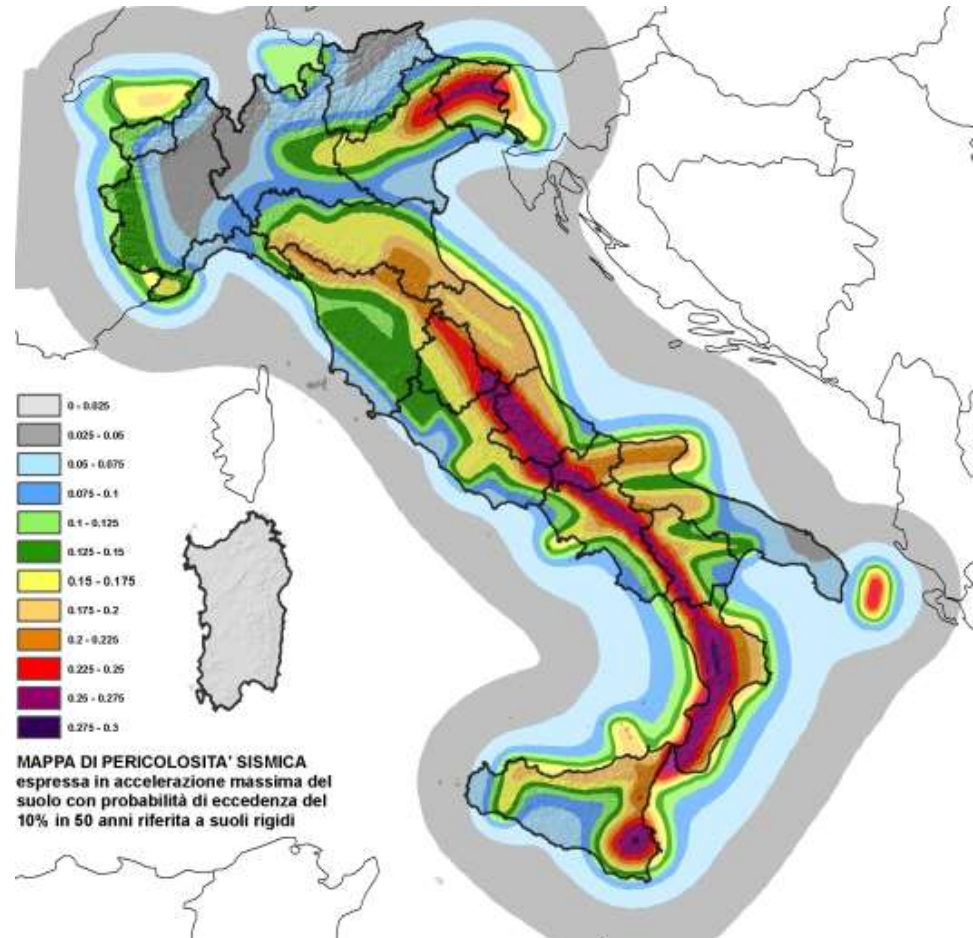
La sismicità dal 1985 al 2014. Sono mostrati i terremoti di magnitudo $ML \geq 2.0$ registrati dalla Rete Sismica Nazionale





Pericolosità sismica del territorio nazionale

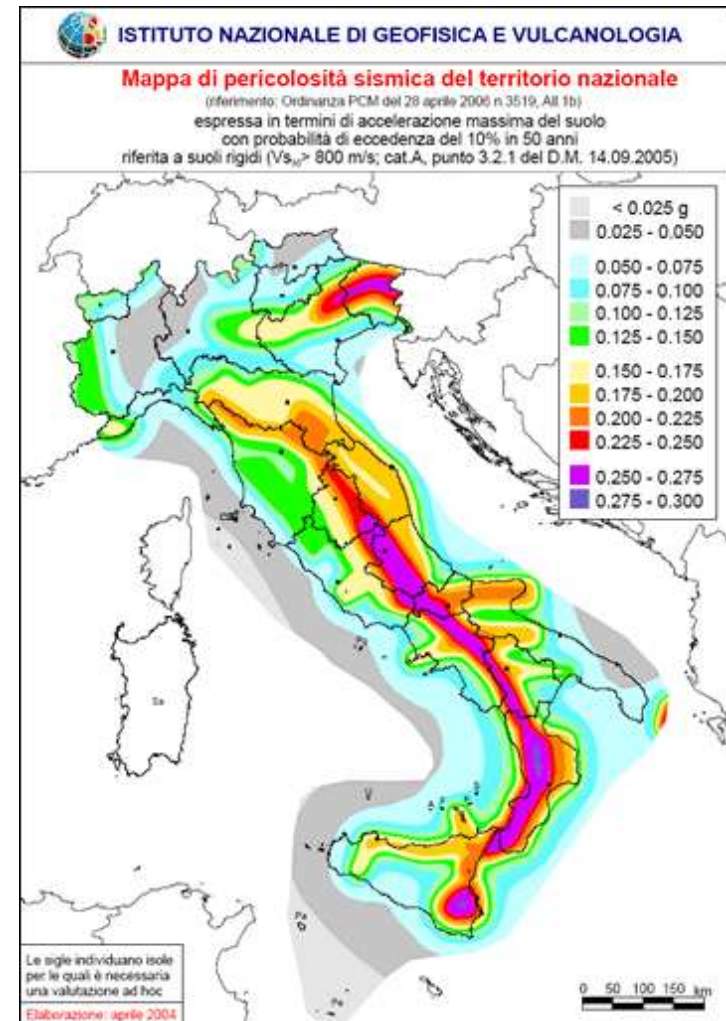
Per questo l'INGV ha realizzato la mappa della **pericolosità sismica** (zonesismiche.mi.ingv.it). Questa mappa si basa sull'analisi dei terremoti del passato, sulle informazioni geologiche disponibili e sulle conoscenze che si hanno sul modo in cui si propagano le onde (e quindi l'energia) dall'ipocentro all'area in esame. Confrontando tutte queste informazioni è possibile ottenere i valori di scuotimento del terreno in un dato luogo a causa di un probabile terremoto, vicino o lontano che sia: tali valori sono espressi in termini di **accelerazione massima orizzontale del suolo** rispetto a **g** (l'accelerazione di gravità). La stima della pericolosità sismica fornisce l'accelerazione massima attesa su suolo rigido con una probabilità di superamento del 10% in 50 anni. Questa mappa, in continuo aggiornamento, mostra la pericolosità delle varie zone dal minimo (colore grigio) al massimo (colore viola).



Pericolosità sismica del territorio nazionale

La mappa di pericolosità è uno strumento fondamentale per la realizzazione di misure di prevenzione che consentano di ridurre gli effetti dei terremoti, per esempio costruendo edifici resistenti alle vibrazioni dei terremoti più forti che possiamo aspettarci in una determinata zona. Sulla base della mappa di pericolosità sismica, infatti, la legge italiana ha classificato il territorio nazionale in 4 zone: dalla zona 1, dove potrebbero verificarsi terremoti molto forti, alla zona 4, a bassa pericolosità ma comunque a rischio per la presenza di edifici vulnerabili.

Circa il 60% dei comuni italiani è classificato nelle prime tre zone. I comuni devono rispettare precise norme sulla progettazione e realizzazione delle costruzioni nuove e sull'adeguamento di quelle vecchie. Tali norme, aggiornate nel luglio 2009, stabiliscono cosa deve essere fatto, in ogni punto del territorio nazionale, in fase di progettazione delle strutture e contengono inoltre nuove regole per il rafforzamento delle strutture esistenti.

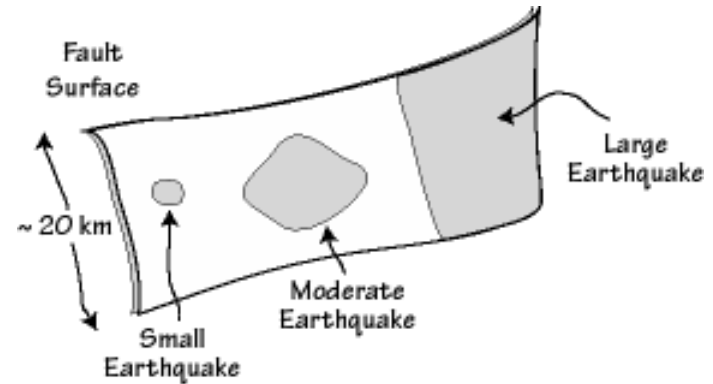




The size of the area that slips during an earthquake increases with earthquake size.

The largest earthquakes generally rupture the entire depth of the fault, which is controlled by temperature.

The temperature increases with depth to a point where the rocks become plastic and no longer store the elastic strain energy necessary to fail suddenly.

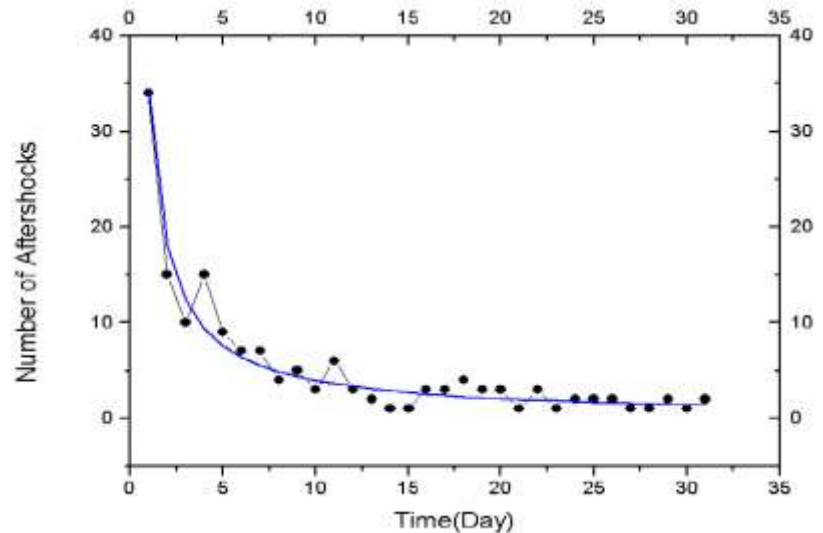
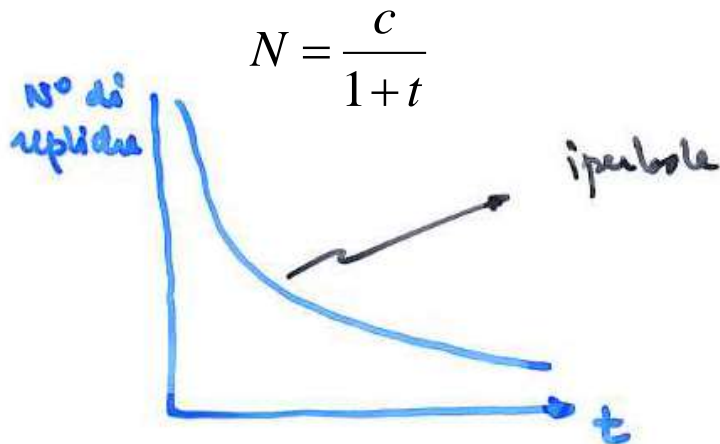


The shaded regions on the fault surface are the areas that rupture during different size events



Earthquake: aftershock

1. Occur after each strong earthquake
2. The biggest replica is typically at least one unit of magnitude smaller than the main event
3. The sum of the energy associated with the sequence of aftershocks is usually 5-10% of that released from the main event → is a secondary process
4. By studying the Japanese earthquakes (1981) OMORI showed that the number of replicas decreases with time after the main event and follows the law of OMORI

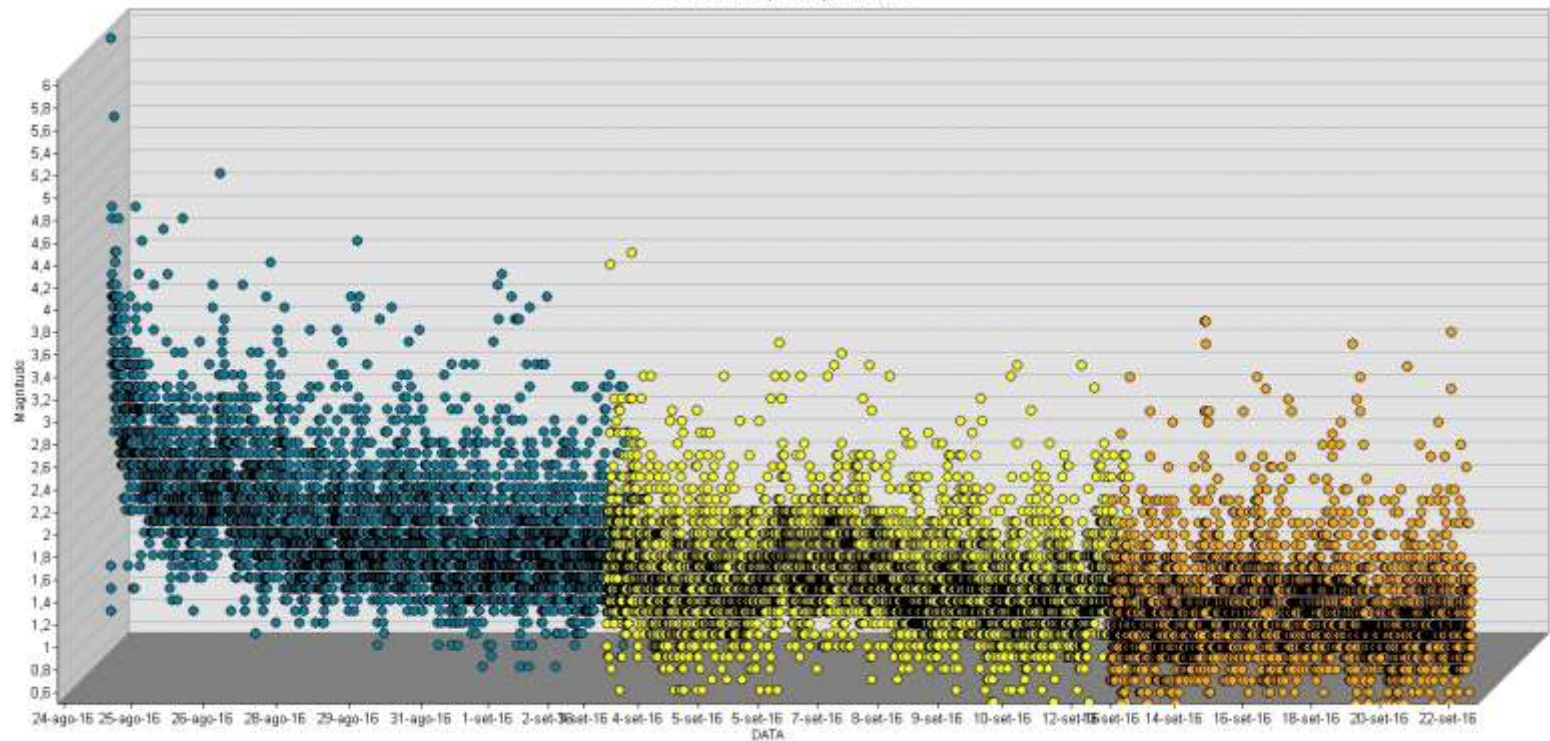




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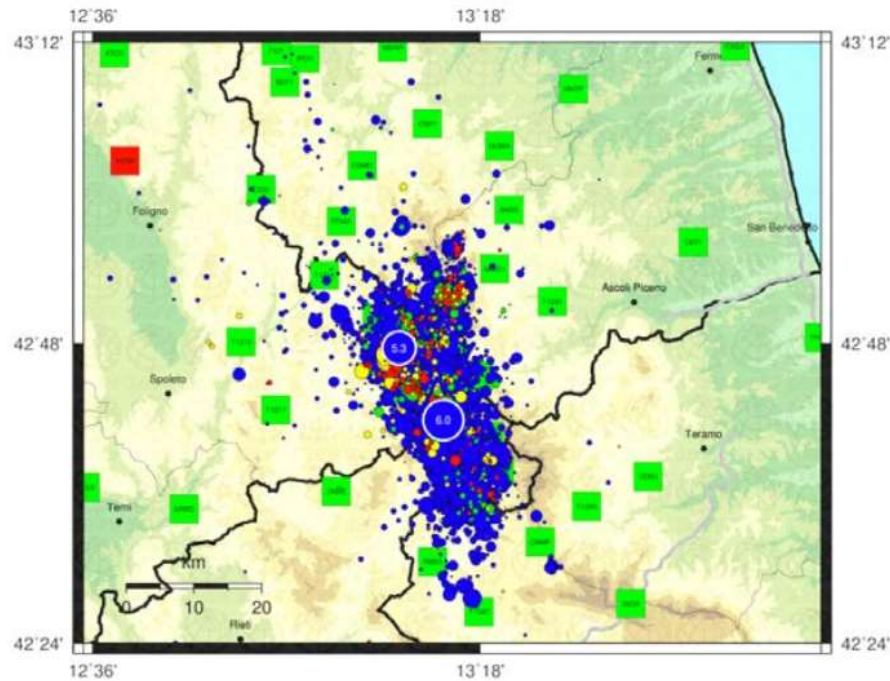
Eventi della sequenza primo mese





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Aggiornata al 2016-09-16, 16:13:05 UTC, numero di eventi 9599

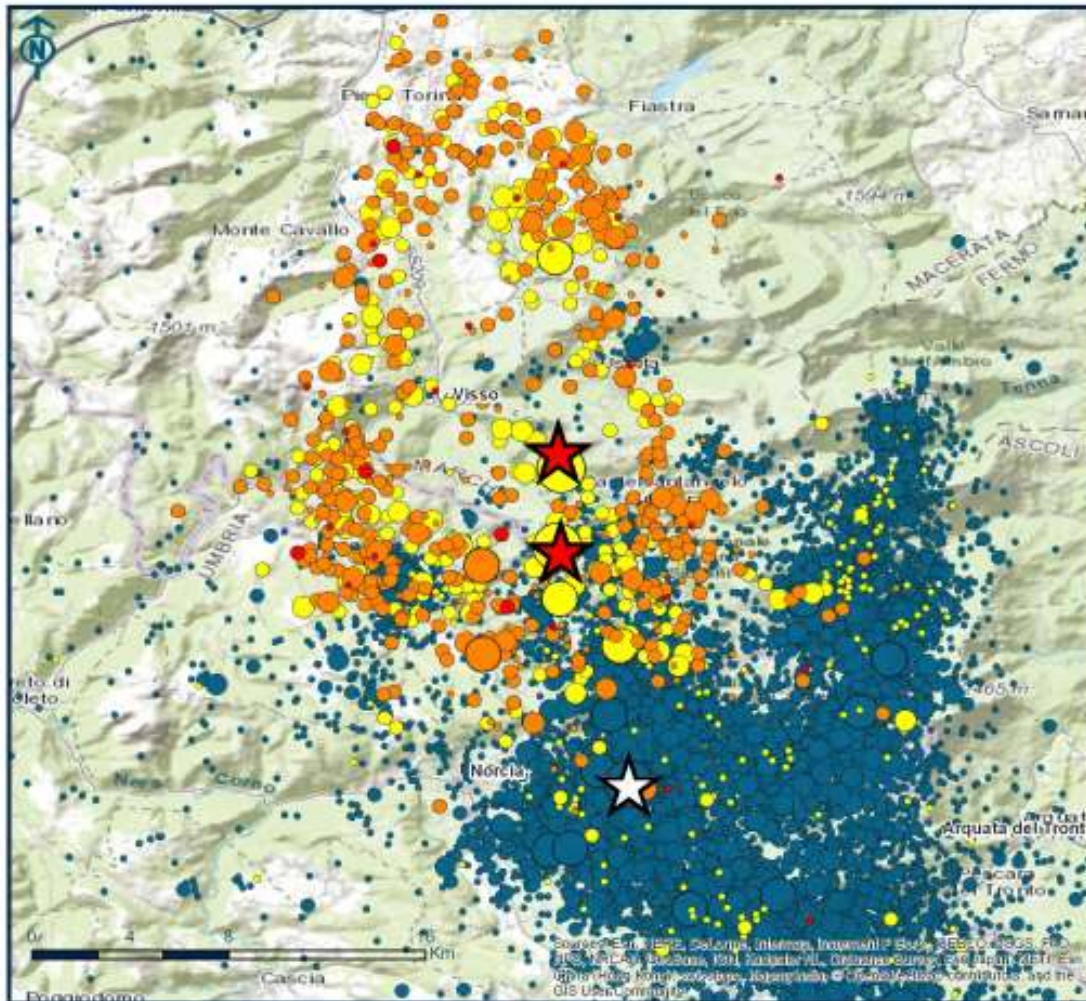
	Oggi	Ieri	2gg fa	Precedenti
MI < 3.0	161	242	309	8687
3.0 <= MI < 4.0	0	7	2	176
4.0 <= MI < 5.0	0	0	0	15
MI >= 5.0	0	0	0	2





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INGV terremoti

**SEQUENZA SISMICA IN
ITALIA CENTRALE
(agg. 28 ottobre ore 10:00)**

Magnitudo (M)

- fino a 2
- da 2.0 a 2.9
- da 3.0 a 3.9
- da 4.0 a 4.9

★ da 5.0

Tempo

- Ultima ora
- Ultime 24 ore
- Ultime 72 ore
- Dal 24 agosto

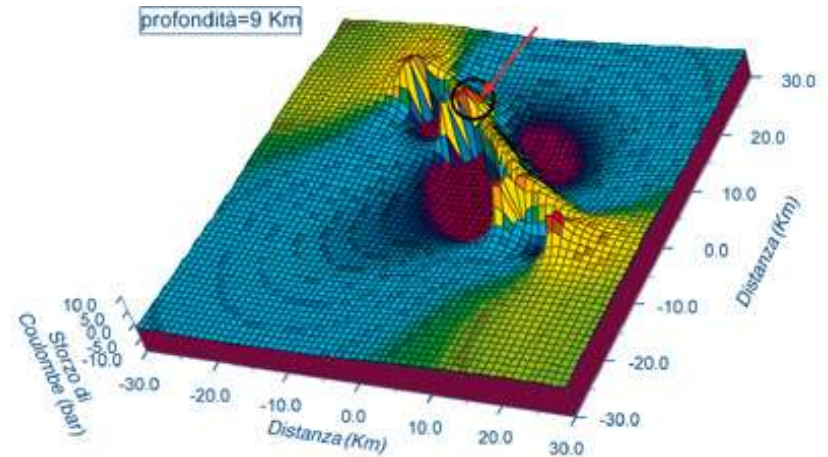
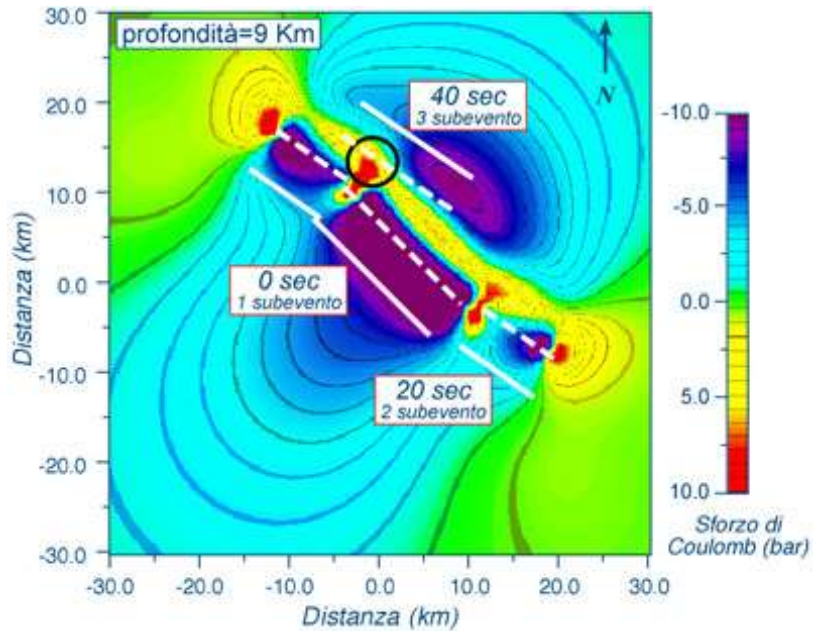
fonte dati: <http://cmt.rm.ingv.it>

SEGUICI SU:
<http://ingvterremoti.wordpress.com>





Terremoto Irpinia 1980





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Visco-elastic modeling and stress evolution since 1511 up to 2004 accounting for coseismic and postseismic deformation

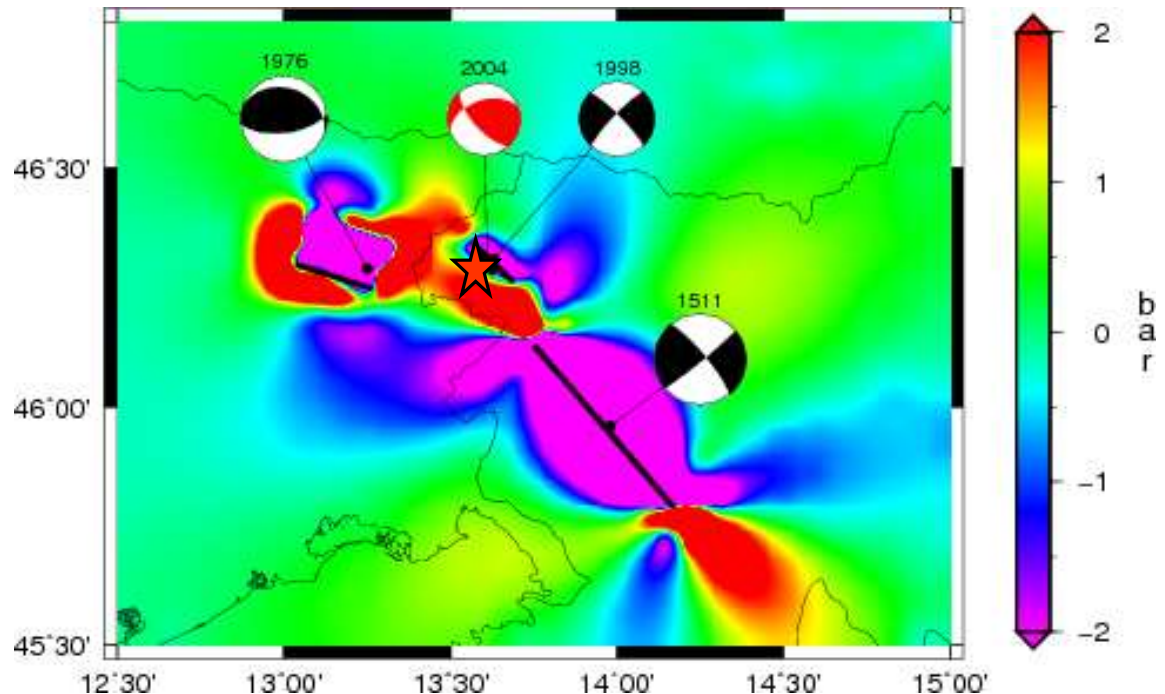
of each past major event

(Borghi, Aoudia- *Tectonophysics*, 2009)

➤ 3-D Finite Elements Method

➤ domain boundaries extending 100 km away from the most external point of each fault

➤ an Earth model comprised by a 16-km-thick elastic upper crust, a viscoelastic lower crust with viscosity 10^{19} Pa s between a depth of 16 km and the Moho at 37 km and a viscoelastic lithospheric mantle with viscosity 10^{21} Pa s.

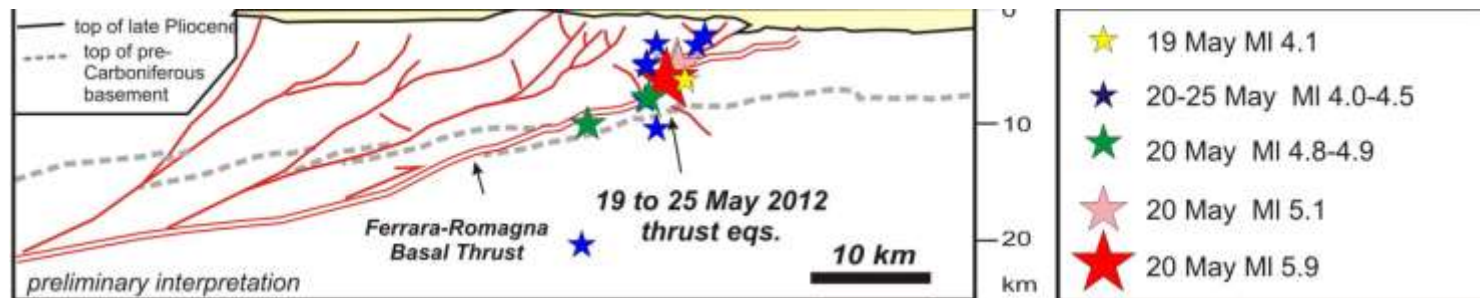


5. The aftershocks occur:

a) On a fault

1. On the fault plane
2. At the end of the fault where stress to increase due to the major earthquake

b) Around the fault



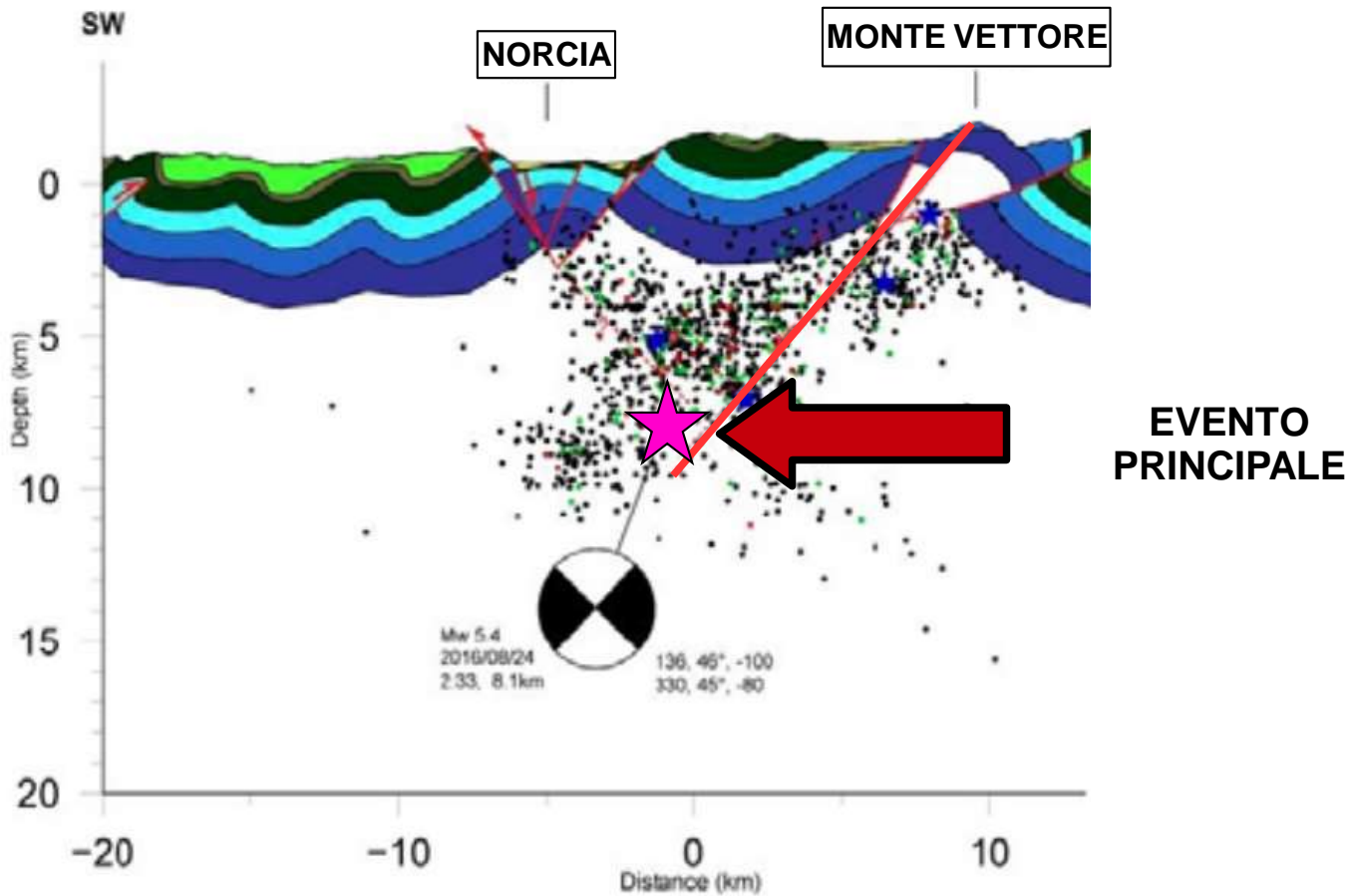
Sezione schematica strutturale dell'area ipocentrale dell'Emilia ritracciata da Toscani et al., 2009 (Da THE FERRARA ARC THRUST EARTHQUAKES OF MAY-JUNE 2012 (NORTHERN ITALY):STRONG-MOTION AND GEOLOGICAL OBSERVATIONS-- REPORT I)



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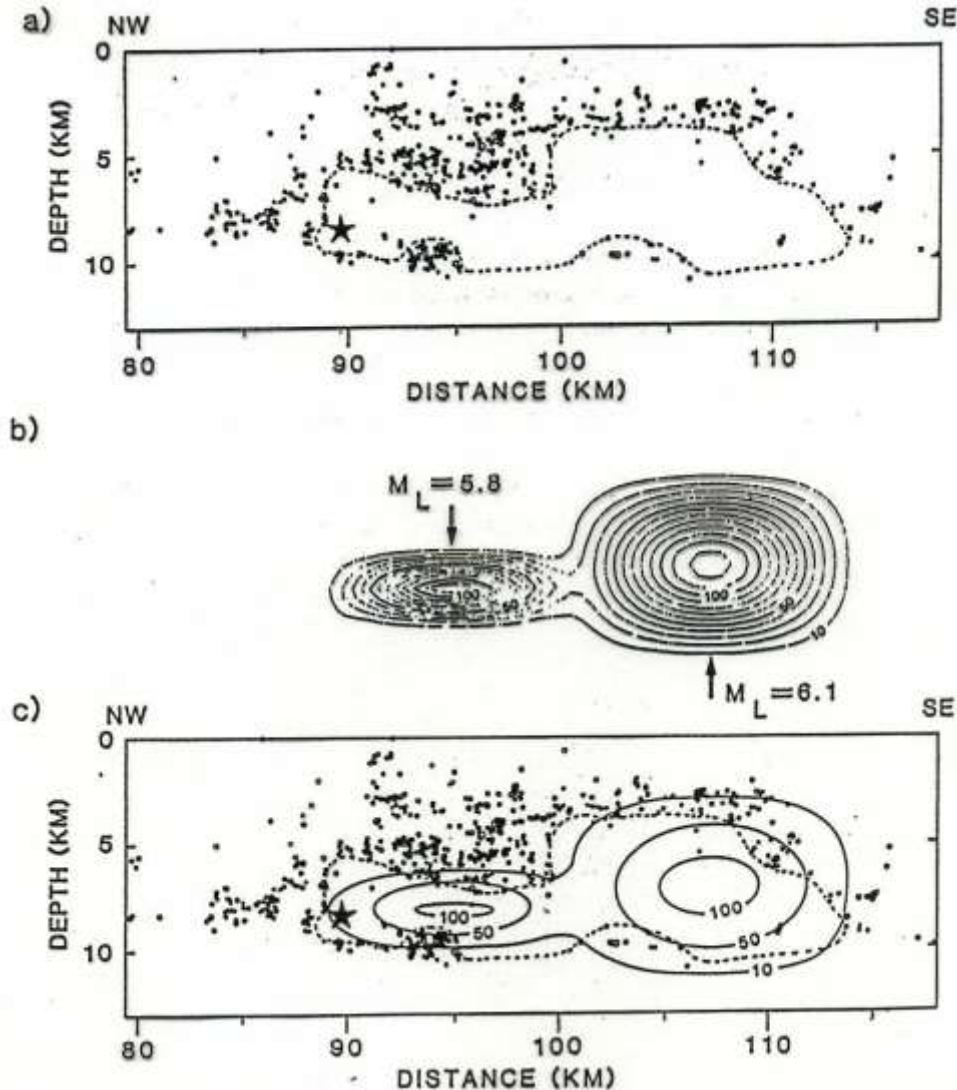
Terremoto Amatrice 24 agosto 2016





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Aftershocks and slip of the earthquake in Morgan Hill (California) in 1984, a strike-slip earthquake on the Hayward let him. (a) the distribution of the aftershocks on the fault, which shows the hypocenter of the main event (star) and the bounds of rupture as reported by Cockerham and Eaton (1984). (b) slip distribution of the two major earthquakes in the sequence. (c) the two overlapping figures (from Bakun, King and Cockerham, 1986).