Magyar Állami Eötvös Loránd Geofizikai Intézet
GEOFIZIKAI KÖZLEMÉNYEK
22

A FÖLDRENGÉSEK ROMBOLÓ HATÁSÁNAK VIZSGÁLATA A FÖLDTANI FELÉPÍTÉS FÜGGVÉNYÉBEN

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Egy földrengés kipattanásakor az azonos távolságban elhelyezett egységes szerkezetű műtárgyak is különböző mértékben károsodhatnak. Ennek néha a mélytektonika, néha azonban a közvetlen környezet különböző földtani felépítése az oka. A különböző földtani szerkezetek gránithoz viszonyított intenzitásnövelő hatása a két-három MSK (MEDVEGYEV, SPONHEUER, KÁRNIK, 1964) fokot is elérheti. Célszerű tehát mérnökszeizmológiai vizsgálatokkal előre meghatározni, hogy milyen lesz a talaj földrengésállékonysága, ha egy területen földrengés következik be.

E vizsgálatokra kidolgoztak néhány módszert, amelyek a talaj rezonáns frekvenciájának, sűrűségének és a benne terjedő szeizmikus hullámok sebességének mérésén alapszanak (KANAI, 1956, MEDVEGYEV, 1962).

E módszerek nehézkesen, de jól használhatók nem túl bonyolult földtani felépítésű területek vizsgálatánál.

A következőkben a talaj földrengésintenzitás-növelő hatásának meghatározására egy egyszerű és viszonylag gyors módszert ismertetünk.

Különböző földtani felépítésű területeken vizsgáltuk a súlyejtéssel gerjesztett szeizmikus hullámok elmozdulásamplitúdóját a gerjesztéstől 20 m-re. Méréseinket különböző energiájú gerjesztéssel — 120 kp, 3 m magasság (*a*, *b*) és 40 kp 1,5 m magasság (*c*, *d*) — többször megismételtük. A mérések elkezdésekor az amplitúdók mellett az első beérkezéseket is feldolgoztuk. Az első beérkezések amplitúdója és az egyes területek földtani jellemzői között összefüggést nem tapasztaltunk. Ezért később csak az észlelt rezgések maximális amplitúdóit vizsgáltuk. A továbbiakban $\bar{A}_x$, $\bar{A}_y$ és $\bar{A}_z$ a különféle talajon mért szeizmikus hullám sugárirányú horizontális, sugárirányra merőleges horizontális és vertikális elmozdulásának maximális amplitúdóját jelenti.

A gránitón mért $\bar{A}_x$ amplitúdó értékét jelöljük $\bar{A}_{xgr}$-rel és képezzük az $\bar{A}_x/\bar{A}_{xgr}$ arányt (1. ábra). A görbék menetét a gerjesztés módja kis mértékben befolyásolja, nagyobb eltérést csak dolomitnál tapasztalhatunk. A horizontális vektor, valamint a teljes elmozdulásvektor maximális értékét a következő módon becsüljük:

$$\bar{A}_H = \sqrt{\bar{A}_x^2 + \bar{A}_y^2}$$

$$\bar{A}_T = \sqrt{\bar{A}_x^2 + \bar{A}_y^2 + \bar{A}_z^2}$$

Képezzük az egyes területekre rendre az $\bar{A}_H/\bar{A}_{Hgr}$, valamint $\bar{A}_T/\bar{A}_{Tgr}$ arányt (2. ábra). A gerjesztés módjától, a vizsgált határokon belül, az amplitúdóarányok

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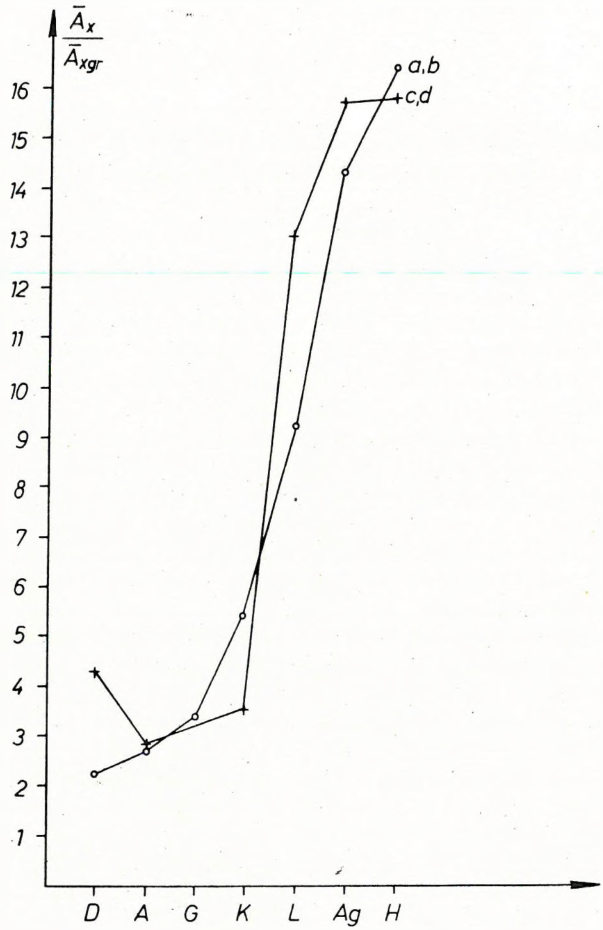
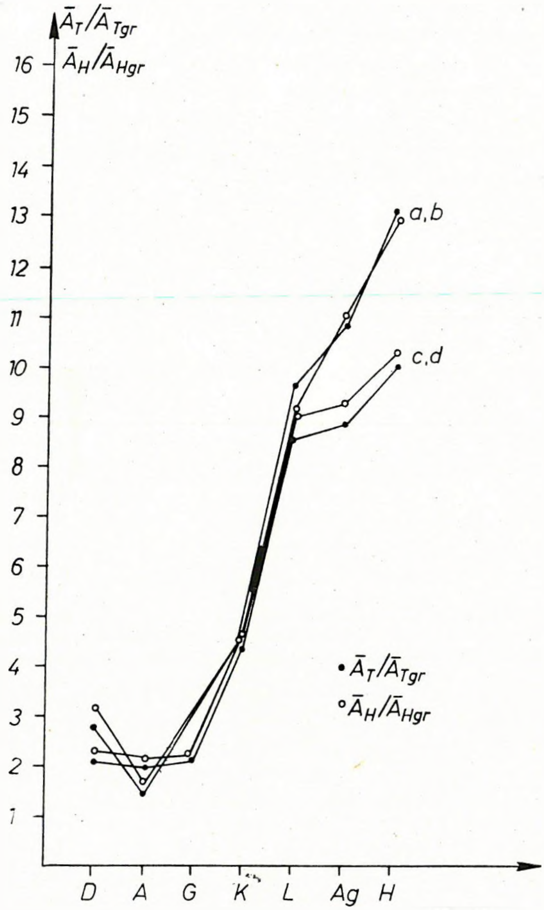
A kézirat beérkezése: 1973. okt. 9.



1. ábra. Különféle földtani felépítés mellett mért maximális sugárirányú amplitúdóarányok összehasonlítása különböző gerjesztésnél *a, b:* 120 kp, 3 m; *c, d:* 40 kp, 1,5 m; *D* = dolomit, *A* = andezit, *G* = gneisz, *K* = dolomitkonglomerátum, *L* = lösz, *Ag* = agyag, *H* = homok

Figure 1 Comparison of radial amplitude ratios for different types of ground and different seismic sources *a, b:* 120 kp, 3 m; *c, d:* 40 kp, 1,5 m; *D* = dolomite; *A* = andesite; *G* = gneiss; *K* = dolomite conglomerate; *L* = loess; *Ag* = clay; *H* = sand

Рис. 1. Сопоставление отношений максимальных амплитуд сейсмических волн, записанных в районах с различным геологическим строением, при различных источниках возбуждения упругих колебаний *a, b* — 120 кг, 3 м; *c, d* — 40 кг, 1,5 м; *D* — доломиты, *A* — андезиты, *G* — гнейссы; *K* — конгломераты доломита; *L* — лёсс; A_g — глины; *H* — пески



2. ábra. Különböző gerjesztésnél mért $\bar{A}_T/\bar{A}_{Tgr}$ és $\bar{A}_H/\bar{A}_{Hgr}$ arányok összehasonlítása

Figure 2 Comparison of the $\bar{A}_T/\bar{A}_{Tgr}$ and $\bar{A}_H/\bar{A}_{Hgr}$ ratios for different seismic sources

Рис. 2. Сопоставление графиков отношений $\bar{A}_T/\bar{A}_{Tgr}$ и $\bar{A}_H/\bar{A}_{Hgr}$ при различных интенсивностях возбуждения упругих колебаний

csak kis mértékben függenek. Medvegyev a különböző talajok intenzitásnövelő hatásának meghatározására az

$$n = 1{,}67 \log V_0\varrho_0/V_n\varrho_n$$

összefüggést találta, ahol

V_n, ill. V_0 a szeizmikus hullám sebessége a vizsgált kőzetben, ill. gránitban, km/sec,
ϱ_n, ill. ϱ_0 a vizsgált talaj, illetve a gránit sűrűsége, gr/cm^3.

Helyettesítsük be $V_0\varrho_0/V_n\varrho_n$ szeizmikus keménységek helyére sorba a $\bar{A}_x/\bar{A}_{xgr}$, $\bar{A}_H/\bar{A}_{Hgr}$ értékeket. Az ilyen módon átalakított egyenletből határozzuk meg az általunk vizsgált területekre az „n" intenzitás értékét.

Az I. táblázatban a kapott eredményeket a Medvegyev által meghatározott „n" értékekkel hasonlítottuk össze. Látható, hogy a homokon mért $\bar{A}_x/\bar{A}_{xgr}$-ból számított kivételével mindegyik intenzitásérték a Medvegyev által megadott intervallumba esik, tehát egyszerű és gyors módszerünk biztató eredményeket ígér. Vizsgálatainkat folytatjuk és ha az eddigi eredményeket bővebb adathalmazzal igazoljuk, a szeizmikus veszélyeztetettség előzetes meghatározása egy könnyen alkalmazható mérnökszeizmológiai módszert kap.

I. táblázat
Table I
Таблица I

Kőzet Rock Горные породы	1,67 log $\bar{A}_z/\bar{A}_{zgr}$	1,67 log $\bar{A}_H/\bar{A}_{Hgr}$	1,67 log $\bar{A}_T/\bar{A}_{Tgr}$	Медведев (1962) n
Gránit Granite граниты	0	0	0	0
Dolomit Dolomite доломиты	0,62	0,80	0,75	0,2–1,1
Andezit Andesite андезиты	0,72	0,3	0,3	0,0–1,1
Gneisz Gneiss гнеиссы	0,80	–	–	0,2–1,1
Dolomitkonglomeratum Dolomite (conglomerate) доломиты (конгломерат)	1,20	1,1	1,1	1,0–1,4
Lösz Loess лёсс	1,60	1,6	1,57	1,2–1,8
Homok Sand пески	2,10	1,67	1,67	1,2–1,8
Agyag Shale глины	1,92	1,63	1,57	1,2–2,1

IRODALOM

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THE DESTRUCTIVE EFFECT OF EARTHQUAKES AS FUNCTION OF GEOLOGIC STRUCTURE

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It has often been experienced that the damages suffered by objects of identical size and in the same distance from the epicentre of an earthquake can be quite different depending on the properties of the ground. In some instances this has to do with deep tectonics, in many cases, however, it is the geology of the immediate neighbourhood which matters. Since the intensity-increasing effect of different geologic patterns as compared to granite can reach as much as 3 MSK degrees (*cf.* MEDVEDEV—SPONHEUER—KÁRNIK, 1964) it has ever been a challenge to engineering seismology to predict the earthquake resistance of different types of grounds.

The various methods proposed (KANAI, 1956, MEDVEDEV, 1962) are based on the measurement of resonant frequency, density and seismic velocity of the ground investigated. A common drawback of these methods is that they are labour consuming and only apply for mildly complex geologic conditions.

The present paper is a summary of our first experiments towards a rapid simple technique of predicting the effect of grounds on the intensity of possible earthquakes.

We measured displacement-amplitudes of seismic waves generated by weight dropping at 20 m from the source on areas of different geology. Measurements were repeated many times with 120 kp, 3 m height (a, b) and 40 kp, 1,5 m height (c, d). As there was no significant correlation between the amplitudes of *first arrivals* and the geologic parameters it seemed appropriate to use the *maximal amplitude* recorded. In what follows let $\bar{A}_x$, $\bar{A}_y$ and $\bar{A}_z$ denote maximal amplitudes of the horizontal, perpendicular and vertical components, respectively. Figure 1 depicts the ratio of amplitude A_x to A_{xgr} i.e. to that measured on granite. The general appearance of the curves is but slightly influenced by the weight applied; the only significant deviation being that for dolomite. The maximum of the horizontal component and that of the total dispalcement can be estimated by

$$\bar{A}_H = \sqrt{\bar{A}_x^2 + \bar{A}_y^2},$$

$$\bar{A}_T = \sqrt{\bar{A}_x^2 + \bar{A}_y^2 + \bar{A}_z^2}.$$

Figure 2 represents plots of the $\bar{A}_H/\bar{A}_{Hgr}$ and $\bar{A}_T/\bar{A}_{Tgr}$ ratios for different areas. Within experimental limits these ratios are independent of the energy of source.

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MEDVEDEV *(op.cit.)* described the intensity-increasing effect of different ground types by the factor.

$$n = 1{,}67 \log V_0\varrho_0/V_n\varrho_n$$

where V_n resp. V_0 (km/sec) are seismic velocities of the rocks investigated and of granite, respectively, and ϱ_n; ϱ_0 are respective densities.

Let us substitute ratio of "seismic solidity" $V_0\varrho_0/V_n\varrho_n$ by A_x/A_{xgr} and A_H/A_{Hgr}; respectively, and determine factor from the modified equation.

In Table I we compare our results with the "n" values determined by MEDVEDEV. With the only exception of the intensity determined from the $\bar{A}_x/\bar{A}_{xgr}$ ratio, for sand, our values fit within the intervals given by MEDVEDEV, so it is hoped that this very simple and rapid technique will prove to be a reasonable substitute of the more elaborate methods published so far for predicting seismic danger spots.

АНАЛИЗ ЗАВИСИМОСТИ РАЗРУШАЮЩЕГО ДЕЙСТВИЯ ЗЕМЛЕТРЯСЕНИЙ ОТ ГЕОЛОГИЧЕСКОГО СТРОЕНИЯ

СЕИДОВИЦ ДЬ.*

При возникновении землетрясения сооружения одинаковой конструкции, находящиеся на аналогичных расстояниях от эпицентра, могут разрушаться в различной мере. В некоторых случаях это связано с глубинной тектоникой, но часто с геологическим строением непосредственной их окрестности. Повышение интенсивности сотрясений за счет различий в геологическом строении может достигать 2—3 баллов МСК (Медведев, Шпонхейер, Карник, 1964). Следовательно, целесообразно проводить инженерно-сейсмологические исследования для определения устойчивости различных видов грунтов против землетрясений.

Для этой цели был разработан ряд методов, в основе которых лежит измерение резонансной частоты, плотности грунта и скорости распространения в нем сейсмических волн (Канаи, 1965, Медведев, 1962).

Эти методы трудоемки, но эффективно применяются в районах с простым геологическим строением.

Ниже дается описание простого и сравнительно быстрого метода для определения сейсмостойкости грунтов. В районах с различным геологическим строением были проведены исследования, причем при помощи падающего груза возбуждались сейсмические волны и на 20 м от источника изучались амплитуды сейсмических волн. Измерения были неоднократно повторены с различной интенсивностью возбуждения — 120 кг с высоты 3 м (*a*, *b*) и 40 кг с высоты 1,5 м (*c*, *d*). Сначала, кроме амплитуд, анализировались и первые вступления. При этом не была обнаружена связь амплитуд первых вступлений с особенностями геологического строения различных районов. Поэтому в дальнейшем изучались только максимальные амплитуды наблюденных волн. В нижеследующем для обозначения максимальных амплитуд горизонтального, перпендикулярного и вертикального смещений сейсмических волн, записанных при различных грунтах, вводятся символы $\bar{A}_x$, $\bar{A}_y$ и $\bar{A}_z$ соответственно.

Пусть величина амплитуды $\bar{A}_x$ в граните будет $\bar{A}_{xgr}$; на рис. 1. представлено отношение амплитуд $\bar{A}_x/\bar{A}_{xgr}$. На основной ход кривых в незначительной мере влияет величина применявшегося груза и только для доломите получается значительное отклонение. Максимальные величины горизонтального вектора и вектора суммарного смещения выражаются соотношениями

$$\bar{A}_H = \sqrt{\bar{A}_x^2 + \bar{A}_y^2}$$

$$\bar{A}_T = \sqrt{\bar{A}_x^2 + \bar{A}_y^2 + \bar{A}_z^2}$$

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На рис. 2. представлены графики отношений $\bar{A}_H/\bar{A}_{Hgr}$ и $\bar{A}_T/\bar{A}_{Tgr}$ для различных районов. В пределах изучаемого диапазона, отношения амплитуд зависят в незначительной мере от интенсивности возбуждения.

Для определения эффекта грунтов, повышающего интенсивность сотрясений Медведевым (1962) было предложено соотшение

$$n = 1{,}67 \log V_0\varrho_0/V_n\varrho_n$$

где V_n и V_0 — скорость распространения сейсмических волн в изучаемых горных породах и в гранитах, соответственно, в км/сек.,

ϱ_n и ϱ_0 — плотность изучаемого грунта и гранита, соответственно, в г/см^3.

Подставим в отношении «сейсмической прочности» $V_0\varrho_0/V_n\varrho_n$ величины A_x/A_{xgr} и A_H/A_{Hgr}. По измененному таким образом уравнению определим величину «n» интенсивности для изучаемых районов.

В таблице 1 дается сопоставление полученных результатов с величинами «n», определенными Медведевым. Из таблицы видно, что за исключением величины, полученной по отношению $\bar{A}_x/\bar{A}_{xgr}$ для песков, все остальные величины интенсивности приходятся в интервал, заданный Медведевым. В связи с этим можно надеяться, что предлагаемый простой и быстрый метод сможет заменить более трудоемкие методы, разработанные до сих пор для предварительного определения сейсмически опасных участков.

EDITORIAL NOTE

A paper of Gy. Dankházi, "Theoretical aspects of the induced polarization method", published in *Geophysical Transactions* XXI. 1–4, 1973, has awakened some comments. The Editor is of the opinion that a too concise composition of the paper which involved neglecting some details deemed as unnecessary by the author has, in fact, rendered the proper understanding of the message difficult.

As a matter of fact, the comments gather around three problems. The author, accordingly, was approached to answer the following three questions.

1. It is known that in actual IP field measurements polarization current is proportional to inducing current. Why has the author chosen just $\varkappa$ (the electric polarization susceptibility) for proportionality factor?
2. In an SI system of units has $\varkappa$ a dimension or not?
3. Why ε_0, the dielectric constant of a vacuum, fails to appear in the exponential expressions?

The author's answers are the following.

To make the point clear it seems appropriate to present a parallel treatment of polarization phenomena in conducting and insulating bodies. Rocks will be considered as conductors or insulators, usually possessing polarization properties. Nonpolarizable insulators behave like *vacuum*, while a nonpolarizable source-free conductor (one for which div $j_0 = 0$) strictly obeying Ohm's law will be termed the *ideal conductor*. In a polarizable rock (insulator or conductor) the electromagnetic field components will differ from those in vacuum or in ideal conductors, precisely because of polarization.

Using the SI system of units in a nonconducting dielectric the change of electric field-strength as compared to vacuum is characterized by the polarization field-strength $\vec{P}_d$ which has been found to be proportional to the field-strength $\vec{E}$ measured in vacuum:

$$\vec{P}_d = \varepsilon_0 \varkappa \vec{E},$$

where ε_0 is the dielectric constant of vacuum in the SI system of units and $\varkappa$ is the electric polarization susceptibility. On the other hand, in case of conducting rocks and other materials it has been found that

$$\vec{j}_p = -\varkappa \vec{j}_0,$$

where $\vec{j}_p$ is the polarization current produced by the inducing current $\vec{j}_0$. The constant of proportionality, $\varkappa$, has been deliberately denoted by the same symbol as the polarization susceptibility figuring in the previous equation even though we realize that there is a considerable discrepancy between notations and opinions of different authors. In what follows we proceed to show the suitability and necessity of this choice of $\varkappa$.

Multiplying both sides of the previous equation by ϱ_0, i.e. the specific resistivity of the rock, we get

$$\varrho_0 \vec{j}_p = -\varkappa \vec{j}_0 \varrho_0 = -\varkappa \vec{E},$$

where we have made use of the equation $\vec{j}_0 \varrho_0 = \vec{E}$. The right-hand side of this equation and of that relating to dielectrics are formally the same but for a minus sign that appears instead of ε_0. Before dwelling upon this any more let us inspect the left-hand side. In the equation for dielectrics, in the SI system of units, the dimensions of $\vec{P}^d$ and $\vec{E}$ differ since ε_0 is a dimensioned quantity. This discrepancy is due to the particular choice of the system of units and can easily be cleared up by making ε_0 dimensionless and attaching the dimension F/m to the relative dielectric constant ε_r. As to the usual alternative, it also has some practical merits since in vacuum the relative dielectric constant ε_r regresses to the dimensionless "1" i.e. to the dielectric constant of the C.G.S. system. Considering, however, the equation multiplied by ϱ_0 we find that all practical advantages disappear in case of *polarizable conductors* when this very important empirical equation yields incompatible dimensions. The contradiction can be settled unambiguously by attaching the dimensions F/m to ε_r instead of ε_0. The dimensions of $\varkappa$ will be F/m as well since the field-strength $\vec{E}$ refers in this case to an ideal conductor for which $\varepsilon_r \vec{E} = \vec{E}$. Here ε_0 is the relative dielectric constant of the ideal conductor, its numerical value being 1. For non-ideal conductors we have

$$\vec{D}_c = \varepsilon_r \vec{E}.$$

Having thus further increased the analogy between dielectric and conducting bodies the polarization of the latter can be described by the equation

$$\vec{P}_c = -\varkappa \vec{E},$$

which is already correct as for dimension.

Let us now investigate the role of the minus sign in the equation for $\vec{P}_c$ and its absence in case of $\vec{P}_d$. Conducting a D.C. to a polarizable conductor we experience that while keeping the voltage fixed the current gradually decreases. This is due to the fact that during the stage of their development dipoles contribute an additional current to that of the free electrons or ions and this additional current dies out by the time dipoles will have been brought about. So, according to the experimental results, polarization current is of opposite direction to the inducing current which accounts for the minus sign. On the other hand, since we cannot conduct a D.C. through dielectric bodies, the current flowing in them is characterized by the displacement current density, i.e. by

$$\varepsilon_0 \varepsilon_r \frac{\partial \vec{E}}{\partial t} = \varepsilon_0 \frac{\partial \vec{E}}{\partial t} + \varkappa \frac{\partial \vec{E}}{\partial t}.$$

This current obviously increases with increasing ε_r bacause of the continuous polarization of dipoles, that is for nonconductors the dielectric constant does not decrease but increases displacement current. The difference in sign between the two equations is a matter of convention and technique of measurement rather than a point of physical significance.

Let us consider now the role of ε_0. The equations for $\vec{P}_d$ and $\vec{P}_c$, respectively, are experimental facts, their validity does not have to be checked. Strictly speaking, these equations hold under certain restricted conditions (homogeneous isotropic rocks, low current densities, etc.) but throughout our investigations we have restricted ourselves to rocks satisfying these ideal conditions, anyway. That ε_0 occurs in one of these equations while fails to appear in the other is an experimental fact and our only task is to try to understand the physical background of this phenomenon. The explanation is pretty straightforward if we approach the question from the point of view of quantum-electrodynamics.

According to quantum-electrodynamics the evolution of dipoles is due to the electromagnetic interaction—transmitted by photons—between dipoles and particles maintaining the electric current i.e. free electrons and ions. Evidently, the interaction between a dipole in a dielectric situated between the plates of a capacitor and the photons moving in the dielectric will be totally different from that between a dipole of a conductor and an ion moving within atomic or molecular distance. The polarization field strength in this latter case is many times higher than that in dielectrics. The difference becomes appreciable if we compare the performance of capacitors made up of dielectrics and of electrolytes, respectively. Capacitors made up of electrolytes can be considered as conductors with $\varkappa \approx 1$ i.e. of an extremely high polarization susceptibility while those comprised of dielectric are insulators. Consequently, in electrolytic capacitors the polarization field strength can be characterized by $\vec{P}_c$ while in dielectric ones by $\vec{P}_d$. Using capacitors of the same size, an electrolytic capacitor yields a capacity of orders of magnitude higher than the dielectric one.

We are now in a position to show why it was necessary to use $\varkappa$ as factor of proportionality in the equation for $\vec{P}_c$. The point is that it is impossible to draw any distinctions depending on the material or rock where polarization takes place. Polarization always results from the interaction of dipoles with the electric field-strength no matter whether this field-strength is due to the voltage between the plates of a capacitor or to ions moving in an electrolyte. There is one and only one physical constant within the scope of quantum-electrodynamics for the description of the relationship between the electric field-strength and the polarized microparticles and this is polarization susceptibility (or the dielectric constant having the same information content).

The introduction of any other physical constants would be justified only if the interaction itself were basically different, e.g. if the development of dipoles were due to mechanic collisions of ions. Classical electrodynamics, however, does not assume nor approve interactions of that kind so if we do not want to violate the idealized conditions and the generally valid range of electrodynamical rules, it is inevitable to use $\varkappa$ as constant of proportionality.

According to what has been said above rocks have to be considered as conducting bodies where an electric field, no matter whether transmitted by ions or brought into existence immediately, can produce polarization. There is no polarization if the relative dielectric constant ε_r is 1 (case of the vacuum or an ideal conductor). Maxwell's equation involving polarization constants becomes

$$\operatorname{curl} \vec{H} = \vec{j}_0 - \vec{j}_p + \varepsilon_0 \frac{\partial \vec{E}}{\partial t} + \frac{\partial \vec{p}}{\partial t},$$

using the same notation as before. The last two terms on the right-hand side correspond to displacement current density. To gain a feeling of its order of magnitude suppose that we perform a field measurement with a current of a frequency $\omega = 100$ Hz and an effective voltage $V = 1000$ V. In this case, if $\varepsilon_r = 10$ F/m,

$$\varepsilon_0 \frac{\partial \vec{E}}{\partial t} + \frac{\partial \vec{p}}{\partial t} = \varepsilon_0 \varepsilon_r \frac{\partial \vec{E}}{\partial t} \approx \varepsilon_0 \varepsilon_r w V = 8.55 \cdot 10^{-11} 10^2 \cdot 10^3 = 8.55 \cdot 10^{-6}.$$

If the resistance of the circuit $\overrightarrow{AB}$ is 1000 ohm, the inducing current is $\frac{U}{R} = 1A$, i.e. the displacement current is only a fraction of some millionths of the inducing current and we are justified in dropping it and putting

$$\text{curl}\, \vec{H} = \vec{j}_0 - \vec{j}_p.$$

This implies that under actual field conditions of the IP method ε_0 does not figure in Maxwell's equations, so it obviously "fails to appear" in their solutions either.

Since the paper under discussion dealt with ideal conductors there had been no need to distinguish $\vec{P}_c$ and $\vec{P}_d$ and the polarization vector and other field characteristics were denoted in the conventional way. It is extremely important to explore the nature of the physical "constant" $\varkappa$ whose constancy, of course, is only one of the idealized conditions we had to accept. In practice, it most likely depends on frequency, temperature and on the texture of the rock. A further simplification we have made is that polarizatin has been attributed to dipoles alone and the role of quadruples and multiples has been neglected.

However, while these and other facts of the theory of IP will remain a task of future research we have to emphasize once more the significance of the quantitative approach, even under idealized conditions as we had done in the paper discussed. The analysis of discrepancies between experimental findings and those predicted by these idealized theories will lead to a deeper understanding of the underlying physics and, after all, to better tools of geophysical exploration.