Milestones in precise Earth observation and their impact

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Introduction

In 1922, the founding year of the German Geophysical Society (Deutsche Geophysikalische Gesellschaft e.V., DGG), the idea of ’continental drift’, advocated by Alfred Wegener since 1912, was still highly controversial. Nowadays, we are used to observe earthquakes and plate tectonics in real time by using recordings of the global seismological networks and observations made with the Global Navigation Satellite System (GNSS). In 100 years of its existence, the DGG has seen a tremendous development of our understanding of Earth’s internal structure and dynamics. This development was only made possible by ever more precise instrumental techniques to measure the globe’s physical properties and changes. Several formerly well established ideas have been questioned and were finally discarded due to the undeniable evidence of new observations.

Much has been written about the history and development of geophysics. It would be presumptuous to try to add relevant new aspects on these few pages and we will not even try to make a complete reference to review literature. Instead we assembled our selective (and likely subjective) timeline in Figure 1, which we complement with anecdotes on certain events. In our small contribution we would like to share with the reader the fascination we feel as we look back on this history of geophysical observations.

Solving the puzzle of global structure

The beginning of global seismology is marked by an observation made by chance by von Rebeur-Paschwitz (1889) with one of his horizontal pendulums, which were actually designed as tiltmeters to monitor the plumb-line at astronomical observatories. Only about 130 years ago, this event took place at a time when little was known about the structure and dynamics of the Earth’s interior and the processes forming the Earth’s surface. Even the (high) speed of wave propagation appeared puzzling at the time (see also supplementary note no. 1: “The speed of seismic waves”). Indeed, neither was the phenomenon of earthquakes well understood apart from being caused by, e.g., *“sudden flexure and constraint of the elastic material”* (Mallet, 1846, see also supplementary note no. 2: “The origin of earthquakes”). It was not before a paper by August Schmidt published in 1888 that seismologists considered the propagation speed of seismic waves to increase with depth and, therefore, pass Earth’s body along curved paths rather than along straight lines. Schmidt (1888) wrote the following about this subject: *“Diese Geradlinigkeit der Erdbebenstrahlen ist eine durch nichts gerechtfertigte Hypothese, welche zwar die Rechnung erleichtert, aber zu sehr zweifelhaften Resultaten führt in der Messung der Fortpflanzungsgeschwindigkeiten und in der Berechnung der Tiefen der Erdbebenzentren, welche ausserdem das Verständnis, die Erklärung einer Reihe von Beobachtungsthatsachen verhindert.”* [1] Such *‘Beobachtungsthatsachen’* (observational facts) were then provided by Oldham in 1900 in the form of the first teleseismic travel times table.

Von Rebeur-Paschwitz soon realized that progress in solving the puzzle of Earth’s internal structure cannot be made from a single observatory, but only with observations collected from all around the globe. In 1895 he wrote: *“Wir wollen in erster Linie die Gründung eines internationalen Netzwerkes von Erdbebenstationen zur Anregung bringen, dessen Aufgabe es sein soll, die Ausbreitung der von grossen Erdbebencentren ausgehenden Bewegungen auf der Erdoberfläche und durch den Erdkörper in systematischer Weise zu beobachten.* [. . .] *Es ist wünschenswert und für den Erfolg des Unternehmens wichtig, dass alle Stationen gleichartige Instrumente wählen und dass diese überall auf den gleichen Grad von Empfindlichkeit gebracht werden.”* [2] That was a very prescient statement and it still applies today (see also supplementary note no. 3: “Global exchange of data”). Global observations are also needed to categorize the energy behind individual earthquakes (see also supplementary note no. 4: “The energy of earthquakes”). Schweitzer (2003) describes

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[1] This hypothesis of the straightness of the earthquake rays is not justified at all. It helps to do calculations; however, it leads to doubtful results in the measurement of the propagation velocities and in the calculation of the depths of the earthquake centers, which in addition prevents the understanding and explanation of a series of observational facts.

[2] In the first place, we want to suggest the foundation of an international network of earthquake stations, the task of which should be to observe in a systematic manner the propagation of the motions starting at great earthquake centers and moving through the Earth’s body. It is desirable and important for the success of this endeavour that all stations choose similar instruments and that these then have the same degree of sensitivity.

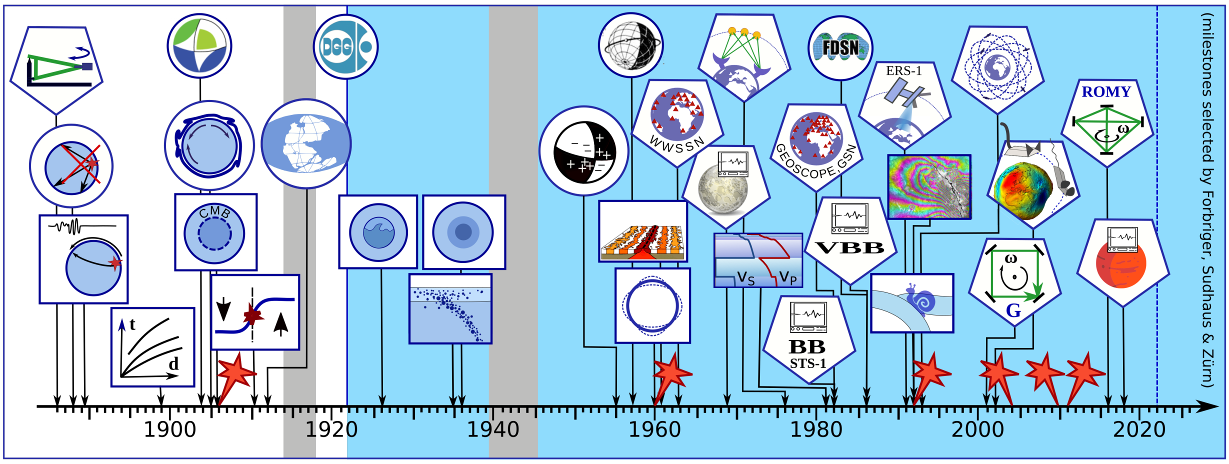


Figure 1: *Selected milestones in geophysics. Boxes: discoveries, circles: new concepts and community events (smaller circles), pentagons: instrumental milestones. 1886 Rebeur-Paschwitz presents his design of a horizontal pendulum, 1888 Schmidt postulates curved seismic rays, 1889 first recording of teleseismic waves discovered by Rebeur-Paschwitz, 1899 Oldham finds body wave phases in travel time tables, 1904 foundation of the International Seismological Association, 1905 Ampferer postulates undercurrents, 1906 Oldham discovers the core mantle boundary, 1906 Great San Francisco earthquake, 1910 Reid finds elastic rebound, 1912 first presentation of Wegener on his continental drift theory, 1922 foundation of the ‘Deutsche Geophysikalische Gesellschaft’ (DGG), 1926 Jeffreys concludes that the Earth is liquid below the core-mantle-boundary, 1935 Wadati-Benioff zones found (Wadati’s paper), 1936 Inge Lehmann discovers the inner core, 1955 Byerly provides fault plane solutions, 1957 International Geophysical Year, 1960 Mw 9.5 Valdivia earthquake (Chile), 1960 first recordings of Earth normal modes, 1961 Dietz introduces the concept of sea floor spreading, 1963 ‘World-Wide Standardized Seismograph Network’ (WWSSN) established, 1969 deployment of the first seismometer on Moon, 1976 first precise ground motion measured by laser tracking of the LAGEOS satellite, 1981–1984 PREM and first global tomographic models of Earth’s structure are presented, 1982 invention of leaf-spring force-balance feedback seismometers (STS-1), 1982 establishment of the global seismometer network GEOSCOPE (1984 GSN and 1993 GEOFON), 1986 the very-broad-band seismograph (Quantagrator) is invented, 1986 founding of the ‘International Federation of Digital Seismograph Networks’ (FDSN), 1991 launch of European SAR satellite ERS-1, 1992 first discovered slow earthquake, 1992 first SAR interferogram of an earthquake, 1993 Global Positioning System (GPS) enters its operational phase, 2001 commissioning of the ringlaser ‘G’ at Wettzell, 2002 start of GRACE satellite mission, 2004 Mw 9.1 Sumatra-Andaman Islands Earthquake, 2010 Mw 8.8 Maule Earthquake, Chile, 2011 Mw 9.1 Great Tohoku-Oki Earthquake, Japan, 2016 commissioning of the four-component ringlaser ROMY in Fürstenfeldbruck, 2018 first deployment of seismometers on Mars’ surface.*

the efforts of Gerland who further pursued the idea after Rebeur-Paschwitz's all too early death. At the end of the 19th century Gerland established an international bureau to collect seismic observations from all around the globe. In April 1904 these efforts resulted in the founding of *The International Seismological Association* (ISA), a predecessor of the IASPEI (*International Association of Seismology and Physics of the Earth's Interior*). Schweitzer (2022) takes this as one of the early roots of the *"Deutsche Geophysikalische Gesellschaft (DGG)"*.

With the availability of more and more seismic records at increasing epicentral distances, the complicated propagation paths of seismic waves could be analyzed to ever greater depths. A landmark is the discovery of the Earth's core by Oldham (1906b) by the conspicuous delay of P- and of S-waves at distances from the source greater than 120°: the core's shadow cast on the surface of the planet. Wiechert had suggested the existence of a core already in 1896 from the differences in mean and surface rock densities. Later, Gutenberg (1913) established the depth of the core-mantle boundary at 2900 km from the travel time curves of P-waves. It took until 1926 when Jeffreys recognized that the core must be liquid because the Earth's rigidity, as determined by tidal deformation, is lower than the rigidity in the mantle, as determined by seismic wave velocities. After explicitly considering the *"elastic stability of the Earth"* he summarizes: *"There seems to be no reason to deny that the earth's metallic core is truly fluid"*. Another ten years later and thirty years after the discovery of the core-mantle boundary, Lehmann (1936) unraveled the complex structure of core-passing wave arrivals and found them to be caused by the Earth's solid inner core.

Seismometers at that time were well capable to record ground motion in terms of a seismogram and not just to indicate the occurrence of shaking. In his introduction to the seismometry of the early days, Wielandt (2002) mentions the names of the builders of the 1930s instruments: Cecchi, Ewing, Rebeur-Paschwitz, Wiechert, Galitzin, De Quervain, and Piccard. The scientist and instrument builder Wiechert himself was one of the founders of the German Geophysical Society, which was called *"Deutsche Seismologische Gesellschaft"* at first and renamed to *"Deutsche Geophysikalische Gesellschaft"* two years later (Koenig, 1974, 2008).

Another milestone in seismology in the early 20th century was the recognition of the rock shear failure process that causes earthquakes, which literally surfaced during the large 1906 San Francisco earthquake and provided the observations for Reid to work out his *'Elastic Rebound Theory'* in 1911 (see also supplementary note no. 2: "The origin of earthquakes"). This earthquake exhibited an unexpected dominant horizontal motion component. Also Alpine scientists already then insisted that significant horizontal crustal movements are needed to explain the shape of rocks in orogenes like the Alps, with Ampferer musing in 1905 on the existence of some kind of *'undercurrents'* below the crust as drivers. Shortly after that, Alfred Wegener (1912) began to propose his concept of continental drift as an alternative earth shape-giving process. For his concept, Wegener assembled many discoveries of different disciplines as kind of jigsaw pieces. Geophysical discoveries have been significant contributions to the concept (see also supplementary note no. 5: "Paleomagnetism revealed moving continents"), and, after decades of dispute, also delivered proof. The final proof of the continental drift is often attributed to Dietz (1961), who proposed the *"spreading sea-floor theory"*. Extensive surveys of the sea floor (bathymetry, magnetism, age, etc.) laid the foundation for this theory. Some of them were carried out in the context of the *'International Geophysical Year'*, which marks a period in the 1950s during which Earth Sciences received many boosts. One of its lasting legacies is the *'Antarctic Treaty'* and continuing international research efforts in Antarctica, though we have to accept that there are not only scientific motivations to keep these projects funded.

Important in the process of verifying Wegener's theory were unexpected observations of significant lateral displacement of crust, e.g., during earthquakes, and puzzling paleomagnetic orientations in rocks (see also supplementary note no. 5: "Paleomagnetism revealed moving continents"). It turned out that seismology and geodesy formed a *dream team* early on in the study of crustal deformation, and their joint application brought great success throughout the history of Earth sciences.

A seismological breakthrough in characterizing earthquake sources en masse was the routine collection of first motions in global teleseismic recording to compile fault plane solutions after a landmark study by Byerly in 1955 (see also supplementary note no. 2: "The origin of earthquakes").

In his report on the concept of sea floor spreading, Dietz (1961) discusses mantle convection as the driving mechanism behind plate tectonics. This concept requires subduction of lithospheric plates into the mantle as the process opposite to sea floor spreading. Evidence for this part of mantle convection was given by the discovery of the so-called *'Wadati-Benioff-zones'* of earthquake hypocenters along downgoing slabs of lithosphere. Wadati reported this in 1935, assembled piecemeal from his papers since 1928 (Suzuki, 2001). Twenty years later Benioff (1955) reported the same discovery independently.

Global exchange of seismic data obviously was a success in that the structure of Earth's deep interior became more and more discernible. Before 1960, seismological research was done with a heterogeneous combination of seismometers of different type and properties, being operated at sites all around the globe and mostly recording on paper. Postprocessing (filtering of recorded seismograms) was practically impossible at the time and waveforms recorded at different stations could be compared only in a very limited way. Efforts to operate global networks of uniform sensors started in the 1960s, also partly as a consequence of the cold war (see also supplementary note no. 3: “Global exchange of data”). A great success was the WWSSN (*World-Wide Standardized Seismograph Network*, Oliver & Murphy, 1971), which recorded on photographic paper. Now, seismograms on microfiches could be obtained by scientists all around the globe, distributed by letter post. The first usable records of Earth’s normal modes were made during the Mw 9.5 Great Chilean earthquake in 1960. Dahlen & Tromp (1998, section ’1.2 Dawn of the Observational Era’) give a very lively written report of how this event opened a new era. As a matter of fact, several strainmeters, the often overlooked non-inertial seismometers, essentially contributed to these observations. Back then, however, recordings still had to be digitized manually to allow a spectral analysis. This called for digital recording but also sparked other instrumental developments. The first digital broadband seismograph was operated at CALTECH as early as 1962 (Miller, 1963) with the intention to preserve the greatest spectrum, dynamic range, and sensitivity. International networks with digital recorders followed. Noteworthy are the *‘High Gain Long Period’* network (HGLP, Savino et al., 1972), the *‘Seismic Research Observatories*’ (SRO, Peterson et al., 1976), and the *‘International Deployment of Accelerometers’* (IDA, Agnew et al., 1976). The latter used LaCoste-Romberg gravimeters with electrostatic feedback and a well-matched 12-bit analog-digital-converter (see also supplementary note no. 6: “The IDA network”).

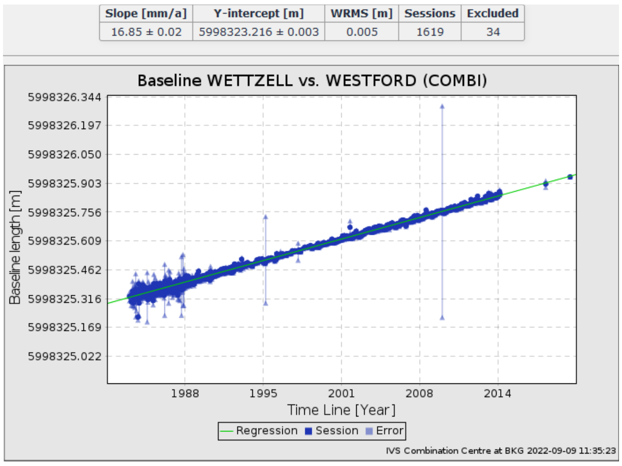
In 1969 Apollo 11 brought the first seismometer to the surface of the Moon. On July 20, 1976, the two Mars landers of the Viking mission had short-period seismometers on board with one of the two functioning (for more details on both missions see the contribution by Knapmeyer et al., 2022). However, no Marsquakes could be identified during this mission due to strong noise caused by the atmosphere. Both space missions mark the beginning of planetary seismology. The advancements of space technology, however, also provided new means to investigate Earth.

Relatively soon after the launch of the first satellite in orbit, *‘Sputnik’*, in 1957, satellites were used for geodetic measurements. Their reference system is the center of mass and therefore independent of moving and deforming tectonic plates. The use of satellites critically complements ground-based sensors in geophysics, e.g., to observe both slow and fast surface movements. The first precise measurement of a position change from space and across a fault was made between 1972 and 1976 by laser tracking of the LAGEOS passive satellite (Smith et al., 1979, see also supplementary note no. 7: “Pioneer measurements using satellites”).

In 1974, when the DGG was about half its present age, the global information on plate motion was gathered and used by Minster et al.: *“Assuming lithospheric plates to be rigid, we systematically invert 68 spreading rates, 62 fracture zone trends and 106 earthquake slip vectors simultaneously to obtain a self-consistent model of instantaneous relative motion for eleven major plates.”* On the distinction of fast and slow tectonic faulting, Kanamori (1977a) provided estimates on the proportions of seismic versus aseismic fault slip. He studied the energy of large earthquakes in Japan using broad-band seismic recordings and put these into context with the measured deformation. He noted: *“We may conclude that approximately 3/4 of the total slip must be taken up by aseismic slip, if the plate motion is uniform on this time scale.”* Today, we frame these processes under the term *‘earthquake deformation cycle’*, coined by Thatcher & Rundle in 1979.

In Germany, space geodetic techniques were taken up in 1970 when the Technical University of Munich and the Institute for Applied Geodesy in Frankfurt started the still very important Geodetic Observatory *‘Wettzell’* in the Bavarian Forest in order to employ space techniques for geodesy in an area which was very dark at all electromagnetic wavelengths. In 1983, the observatory turned into a *‘Fundamental Station for Geodesy’* because essentially all geodetic space techniques were operating there. Today, also many geophysical measurements are carried out in Wettzell. Routine geodetic techniques still include the described laser ranging to satellites, which is mostly used today to control their orbit, but also for laser ranging to the Moon. With respect to plate motion the *‘Very Long Baseline Interferometry’* (VLBI) is a remarkable space geodetic technique with three antennas operating at Wettzell now. Interferometric combinations of quasar signals from far apart ground stations allow for impressively accurate measurements of distance changes between continents which are due to plate motion. Measurement accuracies were improved quickly from decimeters in the early 1970s to below a centimeter today (Fig. 2). As an example, the opening of the Atlantic Ocean can be shown to have a very constant velocity in the last decades, as the remarkably linear increase of the baseline between Wettzell and Westford, Massachussets U.S., of 1.685 cm/year ± 0.002 cm/year shows.

The amount of data collected by global seismological networks being available by the early 1980s enabled the development of global models of Earth’s structure in an unprecedented way. Dziewonski & Anderson (1981) presented the *‘Preliminary Reference Earth Model’* (PREM) from the analysis of normal mode periods and Q values (quality factor that relates to wave attenuation) as well as travel-time data from the ISC-bulletin (*‘International Seismological Center’*). This model has proved very successful in that it is still in use as a reference about 40 years later. At the same time early models of 3D mantle-structure were developed. Masters et al. (1982) used normal-mode data from the IDA network to constrain properties of aspherical structure of the upper mantle. Woodhouse & Dziewonski



Slope [mm/a]	Y-intercept [m]	WRMS [m]	Sessions	Excluded
16.85 ± 0.02	5998323.216 ± 0.003	0.005	1619	34

Figure 2: *Length increase of the baseline between Wettzell and Westford (Massachusetts) as a function of time as determined by VLBI. The slope of this linear trend is* 1.685 cm/year ± 0.002 cm/year *and demonstrates the opening of the Atlantic Ocean by plate tectonics very clearly. The scatter around the straight line as a function of time demonstrates the improvement of the accuracy over time. Courtesy of ‘IVS Combination Center at BKG’ (2022).*

(1984) write in their introduction: *“The accumulation of digital data from the global networks, International Deployment of Accelerometers (IDA) and Global Digital Seismograph Network (GDSN), has only now made it possible to take a global perspective and to begin to construct reliable three-dimensional representations of earth structure, independently of an assumed regionalization.”* Their tomographic model of the upper mantle was accompanied by a similar investigation for the lower mantle (Dziewonski, 1984).

In the 1970s the rapid development of semiconductor electronics provided ’disruptive technology’ as we would call it nowadays. Wielandt (2002) characterizes this period since 1960 as the transition from electromagnetic to electronic seismographs. This finally led to the development of force-balance feedback seismometers, of which the STS-1 by Wielandt (1975) and Wielandt & Streckeisen (1982) is the most famous. It was invented in the early 1980s and turned out to be a true ’game changer’ as we would call it nowadays (see also supplementary note no. 8: “The dynamic range of modern broad-band seismometers”). It was first installed in the *‘Graefenberg Array’* in Germany (Harjes & Seidl, 1978, see also supplementary note no. 9: “At the Graefenberg array”). Soon after, the GEOSCOPE network (Romanowicz et al., 1984; IPGP & EOST, 1982), which celebrates its 40th anniversary this year (2022), was deployed as a global installation of broad-band feedback seismometers. These sensors were soon upgraded to a natural passband from 2.7 mHz to 20 Hz and provided excellent observations of Earth’s normal modes to frequencies as small as 0.3 mHz. Amazingly, the analog bandwidth and dynamic range (140 dB) provided by these instruments still was difficult to be matched by a recording system at the time. Digital recorders were rather limited to 114 dB, used gain-ranging and recorded on magnetic tape with pulse-code modulation (PCM) as Romanowicz et al. (1984) points out. Only with the invention of the Quantagrator by Wielandt & Steim (1986) and Steim (1986), a digitizer with a genuine resolution of 24 bits was available and at last matching the dynamic range of the available seismometers. In 1986, the *‘Incorporated Research Institutions for Seismology’* (IRIS) established the *‘Global Seismographic Network’* (GSN, Butler et al., 2004) to replace the obsolete, analog WWSSN systems. The need of digital recording called for the next effort to establish global standards in instrumentation and data formats. The *‘International Federation of Digital Seismograph Networks’* (FDSN) was founded in Kiel (Germany) in 1986 after a preparatory meeting at the annual conference of the *‘Deutsche Geophysikalische Gesellschaft’* in Karlsruhe, Germany, held together with a meeting of the *‘International Lithosphere Project’* (Romanowicz, 1990). See also supplementary note no. 3: “Global exchange of data”.

Spaceborne observation of Earth’s deformation made a big leap with the installment of *‘Global Satellite Navigation Systems’* (GNSS), which surpassed the absolute positioning precision that was achieved with tracking of passive satellites. The first of those systems and still in place is the U.S. *‘Global Positioning System’* (GPS, see also supplementary note no. 7: “Pioneer measurements using satellites”), which went into operation in 1993. With the active part of the system being the satellites, accurate positioning became possible with comparably cheap receivers, nowadays available in most mobile phones. As a byproduct, GNSS provides an accurate signal of time which can be received all over the globe, and thus solved most of the timing issues in global seismic networks (see also supplementary note no. 10: “Accurate time keeping”). This time signal is generated by on-board atomic clocks and corrected for effects predicted by special and general relativity theory.

Another very relevant development was the introduction of active *‘Synthetic Aperture Radar’* (SAR) sensors in space. With well tracked orbits, these all-weather sensors enable observing line-of-sight position changes of natural and therefore abundant stable points scattering radar waves back to the sensor. Space-borne SAR allowed observation of centimeter-scale relative motion with a spatial resolution of only a few meters and hundreds of kilometers coverage by two flyovers only. A third flyover is needed if the topography is unknown. With these sensors crustal motion detection could be achieved without any ground infrastructure and in very remote places. Variable vegetation, of course, presents a problem. The first coseismic earthquake displacement map was compiled for the 1992 Landers earthquake in California (Massonnet et al., 1993) based on two images from the European SAR satellite ERS-1, launched the previous year in 1991, by using radar interferometry. Now, under certain ground conditions, shallow earthquakes could literally be mapped from space. Radar interferometry is extremely useful also for monitoring changes at volcanoes (e.g. Hort et al., 2022), glaciers and landslides.

In 1995, the first report on the observation of an ultraslow earthquake in 1992, visible in strain meters and not seismometers, was published by Kawasaki et al. (1995). The 800 m long strainmeters at the *‘Piñon Flat Observatory’* (PFO) in California were as well very successful in measuring coseismic strain steps from many medium-size earthquakes and also in identifying the strains caused by plate motion in Southern California (Agnew, 2007; Agnew & Wyatt, 2014).

Seismic hazard assessment and early warning (Lauterjung et al., 2022) have a high impact in society. Though, forecasting the where and when of future large earthquakes is a long-time challenge in geophysics and very likely remains one (Hough, 2009). We are, however, closing in on the ‘where’. Segall & Davis in 1997 observed that *“Detection of slow interseismic strain accumulation is probably the best technique we have for identifying the location of future earthquakes in some areas, because elastic rebound requires elastic strain accumulation prior to earthquakes.”* Slow tectonic motion is estimated based on time series of position changes, for example from InSAR and GNSS. Very recently, compilations of continental-scale, high-resolution strain maps based on time series of SAR interferometry and GNSS measurements emerge, e.g., the relative motion of the Anatolian plate with respect to Eurasia by Weiss et al. (2020) or the motion of the Tibetan Plateau by Ou et al. (2022). These maps enable us to quantify the concentration of tectonic strain at plate boundaries and large fault zones. One of the current challenges for seismology is to use GNSS and InSAR techniques to accurately measure the very slow deformation and transient motions, and to understand their role in plate tectonics and seismic hazards.

In the 21st century not only space technologies are boosting our observational capabilities. An intriguing ground sensor development is marked by the commissioning of the large ring laser *‘G’* in Wettzell, Germany, in 2001. This single ground sensor sends laser light around a 4 m by 4 m square of vacuum light tubes and rivals the very expensive network measurements, which radio telescopes and satellite observations use, in its accuracy in detecting relative changes in the Earth’s rotation of 10^{-8}, and it does so at much higher sub-daily frequencies. *‘G’* is sensitive enough to detect the local influence of the slow *‘Chandler Wobble’* of the Earth’s rotation, a nutation with a period of 443 days, during which the rotation axis changes by only about 9 m at the polar piercing points (Schreiber et al., 2011). Since recently, Germany also hosts another large ring laser instrument, ROMY, at the Bavarian observatory Fürstenfeldbruck. The first laser light in this sensor traveled in September 2016. With four large laser rings in the form of triangles embedded in a giant tetrahedral structure with 12 m arm length, ROMY hosts a laser ring array which renders the measurements more self-sufficient. As a large sensor construction, ROMY needs some time of physical settling and for technical adjustments to perform. Gebauer et al. (2020) state “[...] *once all rings are properly drift compensated and with all scale factors fully established, ROMY presents a viable technique for the continuous observation of Earth rotation and polar motion* [...]”. The ROMY instrument can also function as a rotational seismometer, providing three rotational components in addition to the common three translational ones. Six-component recordings capture all degrees of freedom of rigid body motion and enable new ways of analyzing single-station seismic records for improved earthquake source analysis and earth structure studies (Igel et al., 2021).

After the first observation of Earth’s normal modes in the signals of the 1960 Chile earthquake with moment magnitude Mw 9.5 and the 1964 Alaska earthquake with moment magnitude Mw 9.2, it took almost 40 years until seismologists in Japan realized that the Earth is resonating continuously with its spheroidal normal modes at an amplitude of less than one Nanogal (see also supplementary note no. 11: “Background free oscillations”). It took more than 40 years until the Earth witnessed the next megathrust earthquakes. The Mw=9.1 Sumatra-Andaman Islands Earthquake of 2004 was the first of three megathrust events in the early 21st century. The others are the 2010 Mw=8.8 Maule Earthquake and the 2011 Mw=9.1 Tohoku-Oki Earthquake. Unlike in the 1960s, seismometers and gravimeters of unprecedented sensitivity were now installed around the world. Krüger & Ohrnberger (2005) were able to track the primary P-waves radiated from along the more than 1000 km long fault plane as the rupture progressed. The change of Earth’s gravity field due to the mass displacement going along with the rupture could be measured by the GRACE satellite constellation (Han et al., 2006). After the Tohoku-Oki earthquake, sudden changes of Earth’s density due to P-waves gave rise to newly observed *‘prompt elasto gravity signals’* (PEGS) which propagate at the speed of light, presenting the earliest indication of the megathrust earthquakes at larger distances (Montagner et al., 2016; Vallee & Juhel, 2019; Zhang et al., 2020). Due to the exceptional signal-to-noise ratio realized by these large, and often disastrous earthquakes, Häfner & Widmer-Schnidrig (2013) could measure normal-mode frequencies with an accuracy that put constraints on Earth’s density models. For the same reason, the frequency of the fundamental radial normal mode of the Earth ${}_0S_0$ could be measured to be 814.6566 µHz± 2 ppm, the most accurately known seismic property of the globe (see also supplementary note no. 12: “The frequency of ${}_0\mathrm{S}_0$”), and the strain amplitude of this mode of 10^{-11} could be measured with strainmeters (Zürn et al., 2015).

The success of force-balance broad-band seismometers in investigations of Earth’s structure fueled ambitions to broaden the signal bandwidth for planetary seismology. Only recently, more than 40 years after Viking brought the first seismometer to Mars, the *‘InSight’* mission deployed seismometers on Mars’s surface for the first time, one of them being a three-component force-balance feedback seismometer. The investigation of Mars-quakes and Mars’ internal structure by means of seismology has just started (Lognonné et al., 2020; Knapmeyer et al., 2022).

Where do we go from here?

Many significant developments of new observational techniques have not been included in our historical overview of Earth observation techniques. For example, there is the invention of superconducting gravimeters (Prothero Jr. & Goodkind, 1968; Goodkind, 1999; Hinderer et al., 2007) that were able to reduce the long-term drift in gravimetric records by a factor of 1000, and therefore enable us to observe the changes in the centrifugal force due to polar motion. Atomic beam absolute gravimeters are just starting to become commercially available. The hope is that these will allow continuous absolute gravity recordings, which is not possible with the current absolute gravimeters due to mechanical wear. However, the former still have to prove that they can compete with the accuracy of the latter, which is at the level of 1 µGal (an incredible value of 10^{-9} for the relative accuracy).

There is a general trend towards large-N installations that provide a denser and denser spatial sampling on the ground even on continental scales, e.g., like the IRIS Transportable Array (2003) and the European AlpArray (Hetényi et al., 2018). Also the application of *‘Distributed Acoustic Sensing’* (DAS) to optical fibers that may be several tens of kilometers long, might well contribute to this dense sampling in the future (Jousset et al., 2018). Attempts are made to close the gaps of global seismic networks, which still exist in oceanic regions (Sukhovich et al., 2015). Then there is the tremendous development of laser physics as seen in recent years, which lets us expect further progress in laser gyroscopes, which measure rotation independent of inertia. And there are still more techniques that remain unmentioned here.

In the analysis of precise measurements from highly sensitive instruments it becomes more and more obvious that there is no way to deepen our understanding of Earth’s body, the oceans, and the atmosphere as independent entities. We have to understand Earth as a system, as all its parts interact with each other. For example, atmospheric signals limit the observation of the lowest-frequency normal modes of the globe (Forbriger et al., 2021, see also supplementary note no. 13: “The sensitivity of very broad-band seismometers”) and ocean loading limits the information we can extract from tidal gravity parameters (Baker & Bos, 2003). The atmosphere also interferes with the navigation signals sent from GNSS satellites and causes (space) weather-dependent delays. The need to estimate and correct these delays in turn provides very useful meteorological information on the very dynamic atmosphere, which again feed back into other remote sensing techniques. Such estimations are done, for example, at the Ionosphere Monitoring and Prediction Center (`impc.dlr.de`[3]) of DLR (*‘Deutsches Zentrum für Luft- und Raumfahrttechnik’*) and at GFZ Potsdam in the group on Space Geodetic Technique[4] for tropospheric parameters.

In 2022 we are by no means at the stage where the missing jigsaw pieces are numbered. We still have surprising observations that point to a lack of understanding and which challenge some concepts. On the other hand, we also see holes in the jigsaw for which we are in search of the particular piece. One specific example is the Slichter mode, the translational mode of the inner core. This mode is expected to exist because of physical considerations, but has not yet been detected (see also supplementary note no. 14: “The Slichter mode ${}_1S_1$”). Its properties are so poorly known that we would learn a lot about the density contrast and thermal properties of the material at the inner core boundary by its observation. The missing Slichter mode points to a much larger, still unsolved problem, which is the accurate recovery of the Earth’s internal 3D structure. In particular, we still seek to unravel the characteristics of deep mantle material in terms of density and viscosity (Szwillus et al., 2020). These physical properties are the key to understand the driving forces of mantel convection and the mechanisms behind plate tectonics. Let’s not forget that within our solar system only Earth has plate tectonics, and even living here, we do not know yet why that is the case. In this light, Earth science appears to be planetary science with an exotic specimen.

Some of many emerging techniques will be of good service in the future. One of them, which just became computationally feasible in recent years, is *‘full waveform inversion’* (FWI). After almost one hundred years the dream of one of the founders of the Deutsche Geophysikalische Gesellschaft becomes true. Wiechert (1926): *“Es scheint ein erstrebenswertes und wohl erreichbares Ziel der experimentellen Seismik, jede Zacke, jede Welle der Seismogramme zu erklären und für die Entwirrung der Beschaffenheit der Erdrinde dienstbar zu machen.”*[5]

And last but not least, we must not forget that even with a record of about a hundred years of instrumental observation, we have collected no more than a snapshot compared to the period of the Earth’s evolution.

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[3] https://impc.dlr.de/products (2022-09-16)

[4] https://www.gfz-potsdam.de/en/section/space-geodetic-techniques/overview (2022-09-16)

[5] It appears to be a desirable and possibly achievable goal of experimental seismic work to explain each wiggle and wave in the seismograms and to exploit them for deciphering the properties of the Earth’s crust.

The last century brought such a multitude of important geophysical observations and introductions of new techniques that our report, from the start, was destined to fail in depicting our very successful discipline without large and obvious gaps. We congratulate the Deutsche Geophysikalische Gesellschaft on its anniversary. We wish it will continue to serve for a long time to come and be around as a lively and motivating structure and strong supporter for scientists and students, and as a reliable partner for society in Germany and abroad. May it witness as many fascinating discoveries in the future as it did in the past! And the true discoveries will be those which are currently not foreseen.

Supplementary notes

Here we share some of the notes which we took while collecting material to sketch the story of global studies of the Earth in the last century. Some of them are of anecdotic nature, others present examples of how the precision and accuracy of observations were improved in this period of time. The selection of topics reported here is by no means representative for anything. There would be other institutions, studies, seismic networks, researchers and so on, which may claim with the same right to be considered. Such reports would easily fill a book (if not several books). The topics below are selected randomly and are provided just for the readers' entertainment, not for the sake of completeness.

1. The speed of seismic waves in the Earth appeared amazingly large to early researchers. Cargill G. Knott in 1889 sent a short note to *Nature* in which he appreciates the report made by von Rebeur-Paschwitz (1889) just a few weeks before, which we nowadays accept as the world-famous first teleseismic observation. However, he made a small correction to the time zone used by von Rebeur-Paschwitz to estimate the average propagation velocity of seismic waves to be 2142 m/s. He concludes *"This correction increases the velocity of transmission to 3060 metres per second."* and *"We must assume, then, either that large disturbances in the heart of the earth travel with exceptionally high speeds, or that the origin of the disturbance was a considerable distance from Tokio."*

The average propagation velocity, estimated by Knott (1889), is close to the average wave speed of Rayleigh waves which travel along the Earth's surface as we now know. A compressional wave, which passes right through the Earth's center on its way to the antipode, does this journey in only about 20 minutes, its average speed being about 10 620 m/s and its peak wave speed being almost 14 000 m/s at the core-mantle boundary.

2. The origin of earthquakes was an enigma at the turn from 19th to 20th century, but research to solve it accelerated in this time. Some of the large earthquakes that were caught on global seismograms, mostly thrust earthquakes, produced noticeable fault scarps in the epicentral regions, e.g., the 1891 Mino–Owari earthquake (Japan) and the 1897 Assam earthquake (India). Luttman-Johnson & Day (1898) quotes from a note by R. D. Oldham regarding the earthquake's source *"... we find that the cause of the earthquake was the sudden and permanent displacement of not less than 5,000 cubic miles of the solid crust of the earth, and possibly of five times that volume."* On the source of the former earthquake Davison (1901) wrote: *"The preponderance of preliminary earthquakes* [...] *in 1890-1891 point to the previous existence of the originating fault or faults, and to the earthquake being due, not to the formation of a new fracture, as has been suggested, but to the growth of an old fault."* To the larger forces behind earthquakes Omori in August 1906 points out: *"The ultimate causes of great earthquakes are probably to be traced to the cooling and and contraction of the Earth,* [...]*"*.

The Great San Francisco earthquake on April 18, 1906, caused abundant surface rupture along more than 200 km. Campbell observed in June 1906 *"The motion was principally of the horizontal-shearing type, with few apparent evidences of any vertical component"*. Maybe this fact made Oldham sound somewhat hesitant in his statement in June 1906: *"In some way not fully known, though probably it is more or less directly connected with the gradual cooling of the Earth, the Earth's crust is thrown into a state of strain which ultimately grows too great to be borne, and fracture takes place."* In the same year, the Austrian geoscientist Ampferer (1906) wrote at length about the mechanical shortcomings of the cooling and shrinking theory of Earth with respect to folding within mountain belts and discusses how horizontal *'undercurrents'* (below the crust) could explain the tectonic structures in a comprehensive way. He states that *"Durch die Annahme von selbständigen Ausdehnungen des Untergrundes kann diese Art von Faltenerregung auch ohne Zuhilfenahme der allgemeinen Kontraktion sofort als Erscheinung der Unterstromtheorie begriffen werden."*[6] — and with that postulates already a component of seafloor spreading, again proposed much later in 1961 by Dietz (see also supplementary note no. 5: "Paleomagnetism revealed moving continents").

Regarding the nature of fracturing and faults, Reid (1910), who analyzed the fault displacement of the Great San Francisco earthquake 1906 in detail, mused over the gigantic energy behind these fractures and noticed the exponential fall-off of displacement with distance to the fault. He concluded: *"There is no direct evidence that forces brought into play by the general compression of the earth thru cooling or otherwise were involved, for there is no evidence that the surface of the earth was diminish by the fault."* He also

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[6] With the assumption of spontaneous extensions of the subsurface this type of folding can be comprehended immediately without the help of general contraction as a phenomenon of the theory of subcurrents.

followed: *“This suggests that flows below the surface may have been the origin of the forces we have been considering.”* Reid generalized these observations a year later in 1911 in his work on the elastic rebound theory (Reid, 1911). The coupled processes of long-term slow motion and deformation with sudden failure in earthquakes since then form a strong link between seismology and geodesy.

A whole new cosmos for earthquake seismology was opened when the relationship between wave first motions and faulting was discovered, mostly attributed to Byerly (1955). The elastic rebound theory of Reid (1911) gave a concept of formation of observed fault scarps. Since 1905, seismologists noted down the first motions of arriving waves as seen in seismograms, starting with Omori (1905) and then by more and more seismologists around the world. In the following years, stable spatial patterns were recognized between compressional or dilatational first-arrivals and the epicentral regions from which these waves arrived (Byerly, 1960). It was also Byerly in 1955 who, without doubt, connected the faulting type and seismic first motions patterns unequivocally. Here, seismology and geodesy worked hand in hand again to understand the earthquake source and deciphering signals in seismograms. Thanks to seismology, a real breakthrough was achieved in global tectonics based on *fault plane solutions* that could be extracted from global seismic recordings; for out-of-reach places in very remote areas, beneath the oceans, and for all places with smaller earthquakes that do not reveal their mechanism through surface breaks. From these fault plane solutions resulted then also slip vectors with which seismology handed back information on plate motion for the first global plate model by Minster et al. (1974).

3. Global exchange of data is essential to solve the puzzle of Earth’s deep structure. This has been recognized by von Rebeur-Paschwitz (1895) already in the early days. The global networks nowadays operate stations and share data free of charge across almost all political boundaries. Seismology in that sense pioneered a culture, which in other branches of science now is encouraged by specific funding programs for *‘open data’* and *‘international cooperation’*. In that sense, seismology has been a role model for a peaceful cooperation of mankind on this planet.

Since 2008 data of NASA’s Landsat mission archive are open, provided by USGS, under the framing across *‘Imagery for everyone’* stating[7]: *“It is the policy of the USGS to conduct its activities and to make the results of its scientific investigations available in a manner that will best serve the whole public* [...]. *It brings authority to our data and findings and creates long-term credibility.”* In the planning of the Sentinel-1 Radar mission user groups strongly advised ESA to make the data publicly available, with references to seismology. Since 2014 all data of ESA’s Copernicus Sentinel Program are open.

The open sharing of data is all the more remarkable as we can often see military interests behind the developments. The *‘Vela Uniform’* program funded efforts for monitoring nuclear tests during the cold war and provided the necessary resources for the *‘World-Wide Standardized Seismograph Network’* (WWSSN). The *‘Global Positioning System’* (GPS) was developed to improve navigation capabilities for the United States military and it does not appear far-fetched to suspect military interests supporting the funding of remote sensing satellite systems, which can also be used for military reconnaissance.

In the scientific community the need for international data exchange was recognized early on. However, the technical means of doing so have changed dramatically. In the early days of seismology, paper copies had to be made. Visits of scientists to observatories, the locations of recording, were necessary to get access to waveform data. Wood (1921) compiled a detailed catalog of 312 globally distributed stations. At that time, of course, there was no station inventory available on the internet.

The global exchange of tables of arrival times was comparably easy (through letter mail or later by telex). Since the 1960s the ISC collects these tables of arrival times. This huge collection contains data since 1900. Waveform data were still difficult to exchange. The *‘World-Wide Standardized Seismograph Network’* (WWSSN) was formed in the early 1960s, stimulated by the *‘International Geophysical Year’* 1957 and pushed as well by a need of capabilities concerning a nuclear test ban treaty (Oliver & Murphy, 1971). WWSSN used microfiches for the exchange of waveform data which were sent to researchers by letter mail.

During the early days of digital recording tapes still had to be sent by letter mail from recording stations to the data centers, causing a latency of several weeks. This was all the more difficult the more remote the station is (Schlindwein et al., 2022). Tim Ahern (presentation at the 40th anniversary of GEOSCOPE in 2022) reports that Adam Dziewonski and Don Anderson came up with the idea of an international federation of seismological networks and observatories, which they called PLATO (*‘Permanent Large Aperture Terrestrial Observatory’*) in their brainstorming. This resulted in the founding of the *‘International Federation of Digital Seismograph Networks’* (FDSN) in 1986. This took place during the 46th annual conference of the *‘Deutsche Geophysikalische Gesellschaft’* at Karlsruhe from April 8th to 11th, 1986 (Börngen et al., 1997, Tab. 8). Romanowicz (1990) lists: *“1986, April 10, 11: Karlsruhe (Mid Term ILP Symposium): FDSN preparatory meeting. 1986, April 21,30: Kiel (EGS): FDSN founding meeting.”* With the advent of computer networks, the direct access to digital data became possible.

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[7] See *“USGS Information Policies and Instructions”* at https://www.usgs.gov/information-policies-and-instructions (2022-09-09)

In the 1990s, at the Geophysical Institute in Stuttgart a student assistant was employed whose only job was to periodically log into seismic network stations and download new earthquake data. With the beginning of the internet and corresponding protocols, the exchange of digital data by email or web-browser based download became possible. Ocean bottom stations, however, are still not online and probably will never become (Schlindwein et al., 2022). Larger amounts of data still were sent on magnetic tape (QIC or DAT cartridges) by letter mail in the very early 21st century. This all went along with the development of a variety of data formats for seismological time series. Practically, each recording system and analysis software invented its own type of format. Standardized data formats like the *‘Standard for the Exchange of Earthquake Data’* (SEED) did only partly protect us from the ’Babylonian confusion of tongues’ when it comes to data files. Nowadays, we use clients to access servers on the internet running FDSN webservices directly within the analysis software. Seismic data this way are downloaded on demand. Since the data ’come out of the socket’ practically free of charge, the effort that is expended at observatories and in seismological networks to generate high-quality continuous time series with large dynamic range, large bandwidth, and few gaps is often overlooked.

4. The energy of earthquakes was difficult to assess in the early 20th century. Richter introduced his magnitude scale only in 1935. While scientists studied and compared the near-field effects of large earthquakes carefully, they noted that near-field effects are very individual and depend much on the local geology (e.g. Luttman-Johnson & Day, 1898; Davison, 1901). Therefore, they resorted to categorize the strength of seismic waves in recordings at teleseismic distances. Davison (1906) wrote with respect to the waveforms of the Great San Francisco earthquake 1906: *“If we might estimate the intensity of the shock by the maximum range of movement at Birmingham, we should have to regard the San Francisco earthquake as much stronger than the Indian earthquake of April 1905, but inferior to the remarkable Central Asian earthquakes of July 9 and 23, 1905.”* The vagaries of the instrument response were known back then. He added: *“The period of the larger waves approaches, however, so closely to that of the pendulums themselves, that it by no means follows that the range and epoch of the maximum displacement of the instruments correspond with those of the earth’s crust.”* The development of broadband sensors enabled seismologists to assess the energy based on the wave trains in whole rather than single amplitude peaks. The moment magnitude proposed by Kanamori (1977b) properly relates the measure of earthquake strength to the physical parameters of the rupture process. However, assessing the energy of very large earthquakes remains difficult till today. This is reflected by the extensive discussion taking place after the Sumatra-Andaman earthquake in 2004 (summarized by, e.g., Lay et al., 2005; Menke et al., 2006). This is because even the normal mode seismic signal is band-limited and does not capture the full energy released. Special techniques like the W-phase analysis (Duputel et al., 2012) or data from geodetic observations (Kreemer et al., 2006) are helpful in such cases.

5. Paleomagnetism revealed moving continents and thereby contributed an important geophysical piece in the puzzle of Earth’s dynamics. Rocks may preserve magnetization. Early observations of paleomagnetism in continental rocks had a focus on studying the Earth’s magnetic field and the position of the magnetic poles through time. But some findings were puzzling. Irving wrote in 1956: "*The pole positions calculated from the magnetic directions in rocks of the same age but from different continents should be the same; if they are not, relative land movement between the sampling areas must have occurred subsequently.*". A bit earlier that year, Nairn (1956) stated: "*The inevitable conclusion is that continental drift must be accepted if the assumption of the coincidence of the axial dipole field with the rotational axis is accepted.*" But the answer to the still raging question on how such drift is actually accomplished, remained buried under oceans for a bit and finally was worked out in the concept of seafloor spreading by Dietz in 1961.

6. The IDA network was created after the megaquakes in Chile 1960 and Alaska 1964 and the start of terrestrial spectroscopy. Very few instruments worldwide had recorded the normal modes with good quality, among them are the two LCR gravimeters at UCLA (Ness et al., 1961; Slichter, 1967). In addition, in 1970 gravimeter records of two very deep earthquakes at one station, also in California, allowed for the first time to identify higher harmonics. Efforts to improve the observational situation culminated in the *‘International Deployment of Accelerometers’*, the IDA-network (Agnew et al., 1976; Zürn et al., 1991) of LCR gravimeters equipped with electrostatic feedback with its home at La Jolla, USA. This effort was really rewarded when on August 19, 1977, the Mw 8.2 Sumbawa quake struck. A flurry of observational and theoretical papers on normal modes was triggered by this earthquake and great strides forward in terrestrial spectroscopy were made in the following years (e.g. Dahlen & Tromp, 1998, section ’1.2 Dawn of the Observational Era’). However, the shortcomings of the IDA-network also raised their ugly head: they only measured the vertical component and saturation was reached when the first Rayleigh wave trains (carrying important information about sources and the planet) from big quakes arrived at the station. This led to the development of VBB-seismometers shortly afterwards. However, even today superconducting gravimeters are superior at the frequencies of the gravest modes of the Earth, where the atmosphere is the major player in the noise business and successful corrections for its effects can be applied.

7. Pioneer measurements using satellites of position and changes of position were first accomplished in 1976 by regularly tracking the LAGEOS satellite. LAGEOS was a passive satellite at about 6000 km altitude, reflecting laser light sent up from ground stations. In the moment that the reflected light arrives back without a Doppler shift, the laser beam is orthogonal to the satellite orbit. The rate of Doppler-shift change is controlled by the slant range between the satellite and the ground station. Using a network of stations, the orbit of the satellite can be well determined and relative motion of individual ground stations can be estimated.

A relative distance change of Quincy (California) and east of the San Andreas Fault zone, with respect to San Diego on the west side of the fault, was first determined from space to decrease by 9 ± 3 cm per year by Smith et al. (1979). The first global plate motion model predicted a value much less of about 3 cm to 4 cm per year (Minster et al., 1974), and is also closer to the relative plate motion we measure today. While the satellite techniques advance continuously and while having powerful active sensors, the basic principles, e.g., Doppler-shift tracking, remained pretty much the same through time.

Heading up to a functional GPS system in 1993, quite a number of heavy satellites needed to be launched, an experimental period that covered several years took place during which some geodetic sites have been precisely positioned already before a fully functional system existed. Not coincidentally, the first earthquakes – a doublet on November 24, 1987 – for which GPS-measured coseismic surface displacement have been reported, occurred in the U.S., in California and also on-site of a U.S. military base (Larsen et al., 1992).

Initially, in cold war times, GPS was planned to be used for military purposes only. A tragic incidence in 1983 made the U.S. administration at that time change its mind and plan for a civilian branch of use. Flight 007 from Anchorage to Seoul with 269 people on-board lost its track completely, entered Russian air space and got shot down with no survivors. At this time, the first GPS satellites were already orbiting. If the civilian air traffic had access to GPS, even with a much degraded accuracy, such severe navigation mistakes would not have happened.

8. The dynamic range of modern broad-band seismometers is greatly increased by force-balance feedback. This technique separates the components used for small signal detection in the sensor (mechanical pendulum and displacement transducer) from those responsible for signal conversion (feedback driver and force transducer). Hence both can be optimized for their purpose independently, thus solving the dilemma of the astatic pendulum. Force-balance feedback improves 1) bandwidth, 2) dynamic range, 3) linearity of seismic observations, as well as 4) the reliability of the transfer function of the instruments. Since long it is well known to serve these purposes in the two-pan balance.

Figure 3 compares the dynamic range and bandwidth provided by modern force-balance feedback seismometers with the human hearing range. The dynamic range of the best seismometers of this type is about seven orders of magnitude (140 dB, similar to the best human hearing range and 100 dB more than speech audibility of the human ear). Their bandwidth spans about seven decades (a factor of 10 000 more than the human hearing range and a factor of 400 000 more than speech audibility).

No media of analog recording is capable to capture the dynamic range of modern force-balance feedback seismometers. Just for comparison: If a system recording on paper would use an ink-pen of 0.1 mm diameter, the paper would need to be 1 km wide in order to capture a signal range of 140 dB. For a long time, limitations of recording systems required narrow band-pass filters to be applied before recording in order to limit the dynamic range of the signal. WWSSN always had to operate a short-period system along with a long-period system, both adjusted such that the permanent marine microseisms would not consume the available dynamic range. Even the digital data recorders available at the time when force-balance feedback seismometers were invented, were not able to capture this signal range. GEOSCOPE, for example, started with multiple pre-filtered channels being recorded in parallel with limited dynamic range. This was a severe limitation. Only with full dynamic range digital recording the researcher can make his/her own decision regarding the filters to be applied.

9. At the Graefenberg array (GRF) the broad-band STS-1 force-balance seismometers were first installed between 1976 and 1980 instead of the also available long-period narrow-band instruments from Sprengnether. The decision for these new instruments fell only after Seidl proved by numerical filtering that the broad-band data provide at least competitive quality to the WWSSN-LP instruments besides including the so far avoided frequency band of the marine microseisms. The latter was, of course, only possible with large dynamic range digital recording. Incidentally, the decision in favor of the Streckeisen instrument was severely endangered because of a traffic jam on the German Autobahn causing Streckeisen and Wielandt to show up very late for a decisive meeting in Frankfurt.

10. Accurate time keeping is as essential as the quality of seismic sensors, a fact which often is overlooked. For an analysis of globally distributed records, accurate and stable timing is of utmost importance to measure precise travel times and propagation velocities of seismic waves. The correction applied by Knott (1889) to the report of the Japanese time zone by von Rebeur-Paschwitz (1889) is just one example of how tricky this could be (see also supplementary note no. 1: "The speed of seismic waves").

In the early days, when mechanical pendulum clocks were used, time keeping was a task as difficult as seis-

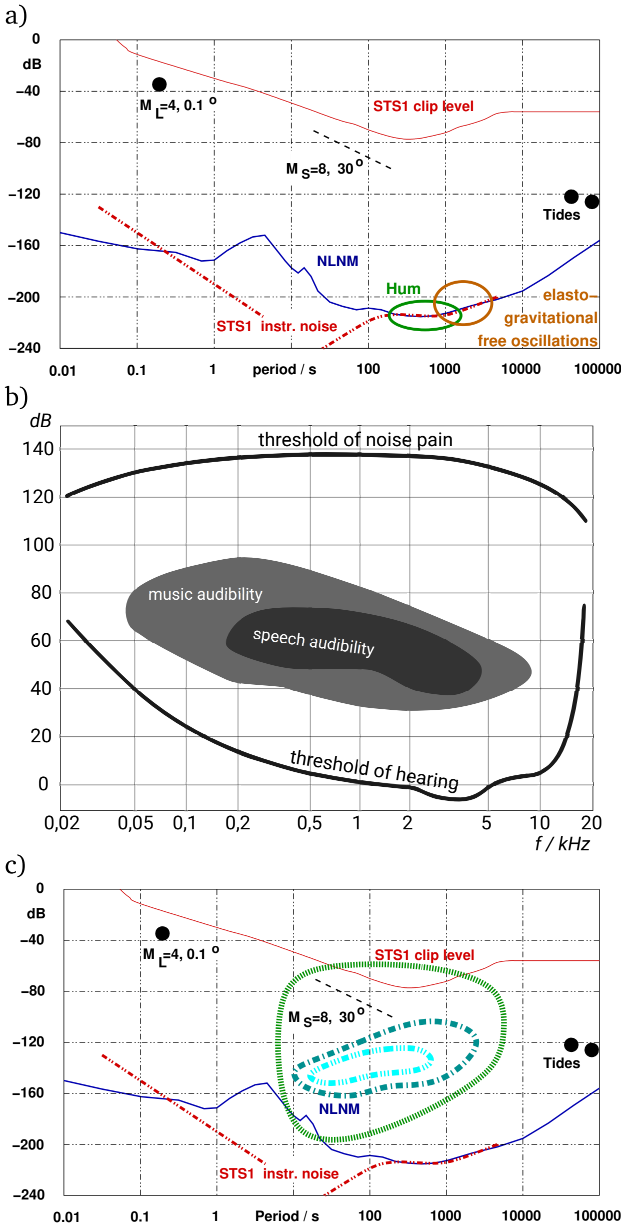

Figure 3: *a): Dynamic range and bandwidth of seismic observations (modified after Wielandt, 2012). Signal amplitudes are given in decibel based on 1 m/s^2. Noise levels are given as RMS-amplitudes in a bandwidth of 1/6 decade. Marks indicate the frequency and signal level of a local and a teleseismic earthquake as well as the tides. The NLNM is the Peterson (1993) low-noise model. It coincides with the level of the background free oscillations (hum). The red lines specify the dynamic range and bandwidth of the STS-1 force-balance feedback seismometer. b): Human hearing range (modified after Hoerflaeche.svg, wikimedia, 2011). c): The human hearing range shifted in frequency and superposed on the diagram of dynamic range and bandwidth for seismic observations.*

mometry. Although they used highly sophisticated mechanics to provide the best available stability, global synchronization was not easy until transmission of electric telegraph signals became available. Still the quality of the reference clock remained an issue. Agnew (2020) dedicates an excellent review to the topic of time keeping in the early days.

When radio broadcasting started, daily or hourly time announcements became available. Later continuous broadcasts were established with one signal per second. In the mid of the 20th century the radio transmission (e.g., by the radio station DCF77 in Germany) of encoded clock signals generated by atomic clocks started and improved the situation. Until seismic stations were equipped with time receivers which were able to decipher the DCF code, leap seconds had to be accounted for manually and were a nuisance for station operators. Nowadays, seismic recorders routinely use timing signals generated from GNSS (*‘Global Satellite Navigation System’*) radio signals. GNSS inherently requires a time signal which is much more accurate than would be needed in seismology. Modern digitizers use a PLL (phase locked loop) to phase lock the sampling clock to the reference time signal which has a jitter at the level of only 1 µs or even less. Radio reception of time signals still is not available for stations at the ocean bottom. This and difficulties at very remote stations prior to GPS being available are reported by Schlindwein et al. (2022).

The timing of seismogram samples is not just essential for the analysis of phase arrivals. In normal-mode analysis the sampling oscillator of the digitizer defines the scale against which the signal frequency is measured. Earth’s normal modes can oscillate several days up to several months (see also supplementary note no. 12: “The frequency of $_0S_0$”) after major earthquakes. *‘Voltage-controlled temperature-compensated crystal oscillators’* (VCTCXO) nowadays easily provide the necessary short- and long-term stability if phase-locked to an external timing signal received by radio (like GNSS or DCF77).

11. Background free oscillations were suspected to be excited continuously by surface processes similar to the excitation of Aeol’s harps by wind. Their frequencies had been computed for realistic models of the Earth already before their actual observation in 1960. This triggered a search for these modes in gravity and strain records by Benioff et al. (1959), because it was expected that the continuously excited resonances should show up in the noise. The authors could only derive an upper limit of their amplitude of about 1 µGal in gravity.

These spheroidal resonances were detected almost 40 years later with much improved instrumentation in 1998 by Nawa et al. (1998), Suda et al. (1998), and Kobayashi & Nishida (1998) at a level of less than 1 nGal in gravity and in the frequency range between

2 and 7 mHz where the noise in vertical acceleration is minimal. Later, Kurrle & Widmer-Schnidrig (2008) detected these continuously excited modes, called *‘the hum’*, also in four horizontal records including now also toroidal modes. The excitation of these modes was found to be caused by oceanic and atmospheric processes at the surface of the planet; however, it is not clear yet why toroidal modes have similar amplitudes as spheroidal ones.

12. The frequency of ${}_0S_0$ is one of the most accuratly measured properties of the Earth. This radial mode has no nodal planes inside the globe and is called the *‘breathing mode’* for this reason. After the Mw >9 Sumatra-Andaman earthquake on December 26th, 2004, ${}_0S_0$ was ringing freely over several months until it received a phase-push by an Mw 8.7 earthquake in the Sumatra (Nias) region on March 28th, 2005. As an example of the determination of the frequency of ${}_0S_0$, Figure 4 shows a phasor walkout analysis for recordings made at the Black Forest Observatory (BFO). ${}_0S_0$ is extremely weakly damped with $Q \approx 5980$. Its amplitude initially was about 1/20 mm and halved about every 17 days. The long time of several months during which the mode was observed, allowed an exceptionally accurate (better than ±2 ppm) measurement of its frequency.

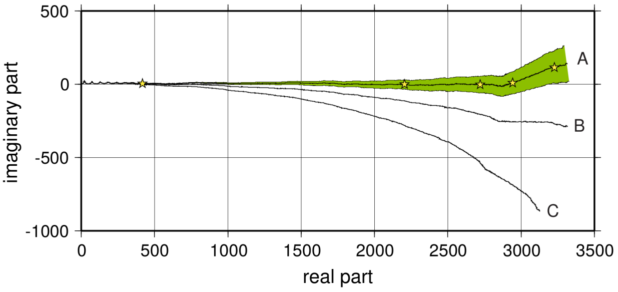

Figure 4: *Phasor-walkout for the radial mode ${}_0S_0$ as observed after the Mw >9 Sumatra Andaman earthquake on December 26th, 2004 from data recorded at the Black Forest Observatory (BFO). The green band indicates a corridor or* ±2 nHz. *A straight line would indicate that the mode oscillates with exactly the test frequency. The phasor-walkout is displayed for three different test frequencies: A:* 0.814 657 mHz, *B:* 0.814 664 mHz±4 ppm *(Riedesel et al., 1980), C:* 0.814 674 mHz ± 11 ppm *(Zürn et al., 1980). The stars mark the dates of January 1st, February 1st, March 1st, April 1st, and May 1st, 2005. On March 28th, 2005 an Mw 8.7 earthquake in the Sumatra (Nias) region kicked the phase of the mode. Zürn & Rydelek (1994) describe the technique of the phasor-walkout analysis. Courtesy of Rudolf Widmer-Schnidrig (pers. comm., 2005).*

The accuracy by which this frequency is known, puts constraints on existing models of Earth’s structure. Standard models like PREM (Dziewonski & Anderson, 1981) do not reproduce the correct value within the error margin. Properties beyond the spherical, isotropic, elastic structure of Earth’s body must be considered. Earth’s rotation and ellipticity changes the frequency by about 336 ppm. Anisotropy may add another 40 ppm and multiplet-multiplet-coupling with neighboring modes affects the frequency at the level of about 50 ppm due to 3D heterogeneity. Anelasticity reduces its value by 5 ppm and without the atmosphere the Earth would ring at a frequency which is 0.5 ppm higher (Zürn & Widmer-Schnidrig, 2007). The latter effect is not significant with respect to the obtained accuracy of 2 ppm. Ding & Shen (2013, their Table 1) list results from recent studies.

13. The sensitivity of very broad-band seismometers under the most favorable conditions allows them to resolve vertical acceleration at the level of the minimum of the low-noise models (Peterson, 1993; Berger et al., 2004), which is near 3 mHz. At lower frequencies the sensitivity of the instruments is such that the level of background signal is dominated by the gravitational signal from mass-fluctuations in the Earth’s atmosphere. Zürn & Widmer (1995) proposed to apply correction procedures commonly used in tidal analysis to reduce the noise level in the normal-mode band. According to Zürn & Wielandt (2007), the minimum in the low-noise models near 3 mHz probably exists because of a cancellation of the involved coupling mechanisms at this frequency. This minimum is at a level of 10^{-12} of gravity. To put 10^{-12} into perspective, let us consider a sewing thread attached to the needle of a well adjusted turn-table record player at its one end. If we then pass the thread over a pulley and attach its other end to an aircraft carrier, the large vessel would experience a horizontal acceleration at the level of 10^{-12} of gravity. This change in acceleration would be detected by the best modern very broad-band seismometers. Sensors operated inside these instruments in order to detect the deflection of the probe mass with respect to the frame are able to resolve displacements at the level of less than 10 pm, smaller than the classical diameter of a hydrogen atom.

14. The Slichter mode ${}_1S_1$ (this name was coined later by Lee Alsop) was tentatively proposed to cause a peak in the spectrum of the gravimeter record of the 1960 Chilean quake at a period of 86 minutes by Slichter (1961). It is basically a translational oscillation of the inner core with respect to the rest of our planet. The major restoring force would be gravitation. However, Slichter needed finite rigidity in the liquid outer core, which was not confirmed by body wave observations, to explain this ”short” period. The Earth’s rotation splits it into a triplet.

Since then the search for this mode (Slichter mode, translational oscillation of the inner core, or ${}_1S_1$ in normal mode speech) continues, but at much longer periods of several hours. The peak at 86 minutes is now thought to have been caused by atmospheric noise. The properties of the Slichter mode would provide unique information about the density contrast at the inner core boundary and other parameters not accessible by other observations. Unfortunately, even megaquakes do not excite this mode to amplitudes observable today, according to theoretical estimates.

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