Daar heb ik veertig jaar over nagedacht...

Feestbundel ter gelegenheid van de 65ste verjaardag van Professor Baarda
De hypothese van de platentektoniek geeft een globale kinematische beschrijving van het aardse oppervlak. In dit model wordt de aardkorst geacht te bestaan uit a-seismische vaste platen die ten opzichte van elkaar bewegen. Nieuwe aardkorst wordt bij oceaanruggen gevormd uit mantelmaterialen en oude aardkorst verdwijnt bij oceaanbotten terug in de mantel. Het model van de platentektoniek geeft echter geen specifieke voor de platenbeweging verantwoordelijke dynamische mechanisme.

Daar de theorie der platentektoniek hoofdzakelijk gebaseerd is op geophysische informatie (o.a. paleomagnetisme en seismische activiteit), geven geodetische technieken de mogelijkheid om de hypothese van de platentektoniek onafhankelijk te testen. Door het verschil tussen de geologische tijdschaal, welke de veronderstelling van vaste platen rechtvaardigt, en de tijdschaal van geodetische metingen, zal men bij toepassing van geodetische technieken tevens de relaties tussen de inter- en intra-plaatbewegingen moeten kunnen modeleren.
Introduction.

One of the most important motivations for precision geodesy is that it provides a means of directly testing the validity of the plate tectonics hypothesis and, at the same time, contributing to the understanding of the many complexities inherent in applying this model to regional and local geological and geophysical observations. The potential of using geodetic measurements to measure the motions of continents has been much discussed and there has been a long history of attempting to measure it by astronomic-geodetic observations. One notable example is Wegener's attempt to measure the drift of Greenland but each time the measurement accuracy was improved the predicted motion was unfortunately reduced. These discussions reached a new height in the late 1960's with the development of two important geodetic measurement techniques, laser ranging to satellites and long baseline radio interferometry. Now, another decade later, it would appear that these goals can at last be achieved were it not for the fact that the realities of plate tectonics are much more complex than the simple model would lead us believe. But possibly we can avoid Wegener's frustrations if some of the geodetic adjustment procedures pioneered by Baarda are employed. These procedures are amply discussed in the preceding papers and it suffices to say that, at least in my memory, Baarda considered the applications of these ideas to satellite geodesy already in 1963. The Palmdale bulge, however, demonstrates that the lesson may not have been learnt.

The majority of papers in this volume will deal with Baarda's theories on geodetic network adjustments and evaluations and I do not wish to dwell on them here. Instead I would like to stress the dynamic nature of the Earth and the geophysics that can be extracted from geodesists' nightmares. The subtle and non-too-subtle
shifts in geodetic benchmarks provide a most valuable boundary condition on geophysical models, always provided that appropriate tests of the geodetic networks have been made.

Evidence abounds for tectonic motions, both vertical and horizontal, on all spatial and temporal time scales. The geological record furnishes many examples of extensive horizontal motions in the form of offsets along major faults while the palaeomagnetic record shows that these motions have occurred globally throughout much of the Earth's past. The geological record also provides abundant evidence for vertical motions in the form of elevated marine deposits or raised marine beaches and terraces. No part of the Earth appears to be free from these motions and deformations. The geologically 'stable' Canadian platform is subject to present-day vertical motions, and Precambrian regions of Central Australia are subject to deformation albeit by small amounts. Neither is the ocean seafloor free from deformation as is seen in the great variety of terrace heights of the islands of the Pacific.

This ubiquitous motion has both a global pattern - as expressed by the plate tectonics hypothesis - and regional and local patterns. Often the latter mask the former and it has only been through the collection of worldwide data that the globally coherent picture of modern-day plate tectonics has emerged. 

The plate tectonics model and geodetic measurements.

The plate tectonics hypothesis represents one of the most important developments in the Earth sciences during the last two decades. Briefly, the hypothesis is that the Earth's surface can be characterized by relatively aseismic units or plates that move relative to each other. New crust is formed at ocean ridges by injection of mantle material and old crust is destroyed at ocean trenches by being subducted back into the mantle. The relative movements of the plates is made possible by a rheological zonation in which the upper tens of kilometers is relatively cold, and brittle and the underlying region is hot and ductile. The upper region is the lithosphere and includes the crust and part of the
upper mantle. The lower region is the asthenosphere. The lithosphere is characterized by its ability to support and transmit non-hydrostatic stresses of the order of a few kbars for long time periods ($\geq 10^6$ years) while the asthenosphere is characterized by its tendency to flow or creep when subjected to even small stress-differences, of the order of a few tens of bars, for time intervals of the order of $10^3$-10$^4$ years.

Observations on which the plate tectonics model rests include the physiographic description of the ocean ridges and trenches, the magnetic evidence for seafloor spreading and apparent polar wander, the distribution of seismic activity and the geophysical and geochemical arguments that go into the model for the rheological zonation. In brief most geophysical observations have gone into building the model with the result that there are few independent observations left to test the fundamental hypothesis. Geodetic observations of the motions between plates or the internal deformations of the plates provide one such test although they do not contribute to an understanding of the mechanism responsible for the motion. Geodetic observations of the Earth's gravity field may provide some constraints on this mechanism but I will restrict the discussion to geometric aspects.

Figure 1 illustrates schematically the spatial and temporal spectrum of tectonic movements. At the low frequency-long wavelength part the spectrum is dominated by the global plate tectonic motions whose average rates are about 5 cm/year but which may attain 10 cm/year or more for the smaller tectonic plates. Also in the low frequency range are a variety of tectonic processes which result in the loading and deformation of the lithosphere. Maximum values may reach 1 cm/year in the case where large volcanic loads are placed in relatively short time intervals of $10^5$ years but more typical values are a few mm/year. On shorter time scales vertical motions result from the recent removal of ice and water loads and rates of a few mm/year and reached in, for example, Fennoscandia, and in the Hudson Bay region. Similar problems occur on a smaller scale, e.g. the uplift of the pluvial Late Pleistocene Lake Bonneville Global.
Figure 1. Characteristic time and length scales of deformations of the Earth.
deforinations of short period are caused by periodic changes in the Earth's rotation and in the lunar and solar attraction. The latter includes the variable and short wavelength tidal loading of the oceanic crust. At the very high frequency part of the spectrum lie the catastrophic deformations associated with earthquakes, expressions of excessive strain having accumulated in the crust. Not all response is instantaneous and frequently relatively slow deformations or creep occur.

The plate tectonics model as outlined very briefly above is essentially a kinematic one in that it does not specify the dynamic mechanism responsible for the motion. Neither does it define the motions in the mantle below the lithosphere. Constructed mainly on geological and palaeomagnetic observations, the hypothesis only represents average motions at the Earth's surface, averages over time spans of the order of $10^5$ years. Studies of the seismicity along the plate boundaries confirm the overall global motions provided that the seismic data is integrated over relatively long time intervals - of the order of 50-100 years or longer. This need to integrate is partly a consequence of the nature of the motion taking place at the boundary and partly a consequence of uncertainties in the data. Nevertheless, significant discrepancies exist in many instances between the average magnetic estimates and the integrated seismic estimates. Along the Marianas trench, for example, the motions are predicted by the plate models to be of the order of 10 cm year$^{-1}$ but no large-moment earthquakes appear to have occurred there during the last few centuries. Near Japan the disparity between the two estimates is a factor of about 5, between Eurasia and India it is a factor of about 3, while along the San Andreas fault system it is perhaps a factor of 2. Such discrepancies at subduction zones may be a consequence of insignificant coupling between the two plates, of an absence of frictional forces, such that the oceanic plate is subducted with minimal 'locking'. Seismic activity will, in this case be much reduced and the seismic method will lead to a gross underestimation of the slip rates. What is required here is a means of measuring deformations at much lower frequencies than is done by conventional instruments.
At the continent-continent boundary between India and Eurasia, the above noted discrepancy may well be a consequence of the deformation occurring over a very wide zone since the large scale tectonics of Asia as a whole appears to be a result of this collision. In this situation it becomes exceedingly difficult to separate the regional trend from more local but related deformations. What is required here is the capability of measuring the spatially averaged deformation over a large area away from the immediate collision zone.

There is little reason to suspect that geodetic observations of global plate motions will not lead to equally difficult-to-interpret results were it not for the ability to measure between points well away from the plate margins. But whether the interiors of plates deform is a different matter to which we will return below. Neither is it obvious that an instantaneous measurement need coincide with the average motions of the past $10^6$ or so years. Also, for a variety of scientific reasons it may be desirable to concentrate on the deformations along or near the plate boundaries since these may throw greatest light onto the mechanism problem.

Seismicity along the plate margins indicate a quite irregular and jerky movement. Very large earthquakes account for most of the energy release and fault slip within a seismic zone and events smaller than "very large" are not significant contributors and are usually considered to be local reactions to the stress redistributions accompanying the larger quakes. If the relative plate motion is regular one would expect the large earthquakes to follow a systematic pattern but the observations indicate that instead the motion occurs mainly in leaps and bounds separated in time by a few decades to a few centuries. Along the Alaska-Aleutian section of the Pacific plate boundary the recurrence interval between large earthquakes is about 60 years while along the Chile section it is about 100 years. At the India-Eurasia boundary the recurrence interval may be well in excess of this. These observations point to long intervals during which a particular section of the plate boundary is relatively stationary or locked by frictional forces. In the presence of a more-or-less steady driving
force acting on the base of the plate as a whole the plate will come under increasing stress near its boundaries. With time the stresses increase until a critical value is reached at which failure occurs reducing stress locally or even regionally. The process may then be repeated elsewhere. In consequence a steady migration of the position of large earthquakes may be anticipated.

Figure 2. A simple model of stress accumulation at a subduction zone. The oceanic plate is under compressions but its motion is locked at the trench. When the stress concentration is sufficiently high a decoupling earthquake occurs at time $t$, decreasing the stresses in the segment of the subduction zone closest to the viewer, but increasing it elsewhere (in the section away from the viewer). This leads to a subsequent earthquake at time $t_2$. $v$ denotes the velocity of the propagation of the stress wave along the subduction zone.
Figure 2 illustrates a possible sequence of events at a continental-oceanic boundary where subduction occurs. At the trench, the lithosphere is held fixed by frictional forces along the interface of the continental and oceanic plates but to the seaward side of the trench there is an upward bulging of the oceanic lithosphere. With time the stresses and deformation increase until one or both of two things occurs. Underthrusting of the oceanic plate may occur if the frictional force at the trench is exceeded and the boundary between the underthrusting plate and the adjacent restraining plate is broken and a temporary decoupling of the converging plates occurs. Secondly the bending moment of the oceanic plate may become excessive, resulting in a tensile fracture in the upper boundary of the lithosphere where the bulge and bending moment are a maximum. This gives rise to two types of major earthquakes; decoupling earthquakes resulting in the underthrusting, and tensile lithospheric earthquakes that break the lithosphere seaward of the trench. Simple plate models suggest that the latter may occur at distances of about 100 km from the trench. The deformation phenomena associated with these earthquakes, other than the obvious co-seismic deformations, are several and include an accelerated plate motion in the vicinity of the boundary before the decoupled lithospheric boundary heals. During this time interval the rate of underthrusting of the down-going slab and the rate of approach of the oceanic plate to the trench will increase while the stress previously built up in the plate as a whole is relieved. On the continental side of the trench a sinking of the lithosphere may occur while on the oceanic side there may be a reduction in the elevation of the lithospheric bulge. The post-seismic deformations may be as high as 1-5 m/year, decreasing with time and are confined to within about 100 km of the boundary. They are a result of stress being released within the interior of the plate and the time taken for this release to be completed may be several decades depending on the viscosity of the underlying asthenosphere, that is on the nature of the coupling between the lithosphere and asthenosphere.

The pre-seismic deformations, obviously of great importance for studying premonitory behaviour, are also informative on the rheology of the crust. Geodetic
evidence for them go back at least as far as the 1920's, but the interpretation of this data has not always been clear, in part a consequence of the complexity of the phenomena and in part a consequence of unsatisfactory surveys.

Whether plates as a whole move continuously or intermittently has been frequently debated. Do plates respond instantaneously to applied forces or not? The answer to this question depends largely on the contrast in viscosity between the lithosphere and asthenosphere. If this contrast is very great there will be little or no drag force at the base of the plate and irregular motion is possible. Normal loading problems indicate that on time scales of about $10^6$ years the asthenosphere can be considered as a fluid while on time scales of about $10^3-10^4$ years this is no longer so. This suggests that 'instantaneous' response here means the average response over periods longer than $10^4$ years but less than $10^6$ years and that irregularities in plate motion can be expected on time scales of greater than about $10^4$ years if changes in driving mechanisms or boundary conditions permitted this.

An important axiom of the plate tectonics hypothesis is that the plates behave as essentially rigid entities. It would be truly remarkable if this is so, if large irregularly shaped surfaces subjected to a variety of driving and resisting forces, can be moved over an ellipsoidally shaped Earth without undergoing some deformation. What permits geologists to make this assumption is that the time scale of their model is quite different from that of the average lifetime of geodesists, the stressing and straining of a continent seen in the instantaneous picture, largely vanish when integrated over long time intervals.

Clearly in relating the instantaneous measurements the relationship between inter and intraplate motions must be well understood. Evidence that continents as a whole are subject to differential stresses is readily seen in the distribution of seismic activity within the continents of North America, or of Australia. Sykes summarized the distribution of the intraplate earthquake and of igneous activity and one of the principal conclusions he reached
was that in continental areas seismic activity tends to be concentrated along pre-existing zones of weakness - along unhealed faults - within areas affected by the youngest orogenesis that predated the opening of the present oceans and led to the present cycle of plate tectonic activity. This seismicity is presumably activated in response to the present-day stress regime in the plate but which is not necessarily the same as that which created the zone of weakness in the first place. Examples abound, eg. the Rhine graben in Europe or the Adelaide Geosyncline in South Australia. Zones of weakness within the plates are usually delineated by seismic activity and can therefore be avoided. In any event, these intraplate deformations will generally be much less than the interplate motions and provided that the stations lie far away from plate boundaries geodetic observations should provide a satisfactory test of the hypothesis. This is assuming that the geodetic methods have already been demonstrated to be satisfactory in themselves. That this is not so emphasizes the need to first carry out careful geodetic experiments within a single plate. Some experiments in this direction have been carried out and these indicate that the accuracy of inter-station baselines remains worse than the measurement accuracy, a sufficient warning that systematic model errors still remain and that the careful testing of any geodetic results is of the utmost importance.

An area where the geodetic measurements will not contribute is in resolving the geophysically vexing problem of absolute motion, a concept that has a different meaning here than in geodetic usage. One day it may be possible for geodesists to detect motions of their tracking stations with centimeter accuracy within an absolute celestial reference frame but in geophysics it is the motion of these stations relative to the inaccessible interior of the Earth that is of physical importance. An overall westward motion of the lithospheric plates over the underlying mantle, for example, would not be detected by the geodetic measurements. One would erroneously interpret such a coherent drift as a change in the speed of rotation of the Earth although it is only the crust that is involved. It is the same as the older problem of the separation of
continental drift from polar wander. Geophysicists have resolved the problem in a seemingly ad-hoc way with the introduction of Wilson's hotspot model and with the associated axiom that the hotspots are fixed relative to each other and to the deep mantle. The result is apparently self-consistent and the model is useful but it cannot be tested by external, geodetic measurements.

Vertical motions, while not playing a central role in the plate tectonics hypothesis, accompany the horizontal deformations, particularly along the boundaries. The elevated areas of some of the collision zones should be ample evidence of this. Other examples are the progressive uplift of parts of New Guinea as witnessed by a series of elevated beaches and marine terraces. An example of the interaction between the horizontal and vertical motions is illustrated by Figure 2. The accumulation of stress at the locked boundary could lead to a flexing of the lithosphere on either side and to uplift. When decoupling of the lithosphere occurs such that stress at the boundary is released one or both sides may be subjected to subsidence. Some well-documented case histories exist which demonstrate the vertical motion during the stress cycle. Differential uplifts of 1 m over some 100 km occurred during the co-seismic phase of the Nankaido earthquake of 1946 in southern Japan. Pre-seismic vertical motions of 10-20 cm were recorded over the same area for some 30 years leading up to the event. Post-seismic deformations over the next two decades were of comparable magnitude.

Within the plate interiors vertical motions in response to variable surface loading conditions are common, although these are not usually associated with earthquake activity. Well-known examples include the post-Pleistocene rebound of Fennoscandia and the Laurentide region where the presently observed uplift rates are of the order of a few mm per year. Other examples include the rebound of desiccated lakes, of which Lake Bonneville is perhaps the best example. Geodetic observations specifying both the regional deformation pattern of beaches or terraces and the present rates of uplift, when taken together with geological or geomorphological observations of vertical motion, have provided useful information on the viscosity of the mantle below the
lithosphere. Viscosity estimates of the mantle below the oceans are more difficult to come by. One possible example is for Iceland where extensive glaciation and subsequent rebound has also occurred but here the problem is complicated by the presence of an active spreading centre which may also be associated with vertical motions. Elsewhere in the oceans subsidence of some islands has also been noted, a few mm/year in the case of the Hawaiian islands for example, and this has been interpreted in terms of a viscous relaxation of the stress in the oceanic lithosphere due to the volcanic load. Island uplift and subsidence may also occur in the vicinity of recent volcanic formation where the lithosphere is flexed upwards by the application of the surface load. Other examples occur when the lithosphere passes over a heat source in the asthenosphere resulting in thermal expansion and uplift. All of these problems are closely linked with problems of variations in sealevel but this is not inextricable.

Conclusions.

In a sense, the geodetic requirements for monitoring the motions and deformations associated with the plate tectonics hypothesis are readily stated: A dense and highly accurate global network of stations whose three dimensional positions can be determined at regular and frequent time intervals. The realities of resources, geography and politics soon makes nonsense of such ideals and each tectonic problem will have to be investigated separately. Important applications are to be made in areas close to and across plate boundaries, at continent-ocean collision zones such as along the Pacific side of South America, along the ocean-ocean collision zones of the Tonga and Kermadec Isles, at the transform faults of Anatolia and New Zealand, and at the continent-continent collision of Eurasia with Africa and India. In many of these instances a combination of terrestrial methods with extra terrestrial measurement techniques could go a long way to understanding the tectonic processes that are a consequence of the global motions.

As hinted at already, I do not necessarily think that the most important aspect of applying geodesy to
geophysics is the testing of the fundamental notion that the tectonic plates move relative to each other. The geophysical and geological evidence that they have done so through much of the past appears overwhelming even though this evidence may not be uniformly accepted - or understood. Attaching such unimportance to testing a basic hypothesis may seem to put me at odds with Baarda's philosophy of always testing the hypotheses upon which one constructs models and theories. But this is not necessarily so. The design and execution of an experiment to measure the motions of the major tectonic units is such a major undertaking that I rather doubt whether a thorough test is possible without the establishment of a long term and global network so that regional and local deformations become separable from the more global trends. In the meantime I would prefer to see the geodetic techniques applied to more specific problems which would contribute to a clearer understanding of the rheology of the crust and mantle, of the mechanism responsible for the motions and of the physics of the tectonic processes.

References.


6. X. Le Pichon, J. Francheteau and J. Bonnin, Plate

7. See, for example, the papers in Glacial Isostasy, edited by J.T. Andrews (Dowden, Hutchinson and Ross), 1974, and Earth Rheology, Isostasy and Eustasy, edited by N.-A. Morner (Wiley), 1980.


15. H. Kanamori and J.J. Cipar (Phys.Earth Planet.Int., 9, 128-36, 1979) observed a slow deformation over a period of about 15 minutes in the epicentral area prior to the major shock of the 1960 Chile event. Other evidence for the 'silent' events is very limited, A.M. Dziewonski and P. Gilbert (Nature, 247, 185-8, 1974) have reported stress release just before two deep events and W. Thatcher (Science, 184, 1283-5) finds aseismic slip associated with the 1906 San Francisco earthquake.


23. See, for example, papers in some of the early volumes of the Bulletin of the Earthquake Institute of Japan.

24. See, for example, some of the conflicting conclusions reached about strain accumulation in Southern California by W. Thatcher (Science, 194, 691-695, 1976 and loc.cit. note 26) and J.C. Savage and W.H. Prescott (J.Geophys.Res., 84, 171-177, 1979).


