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PLATE TECTONICS

XAVIER LE PICHON, JEAN FRANCHETEAU and JEAN BONNIN

Centre National pour l'Exploitation des Océans,
Centre Océanologique de Bretagne, Plouzané (France)

ELSEVIER SCIENTIFIC PUBLISHING COMPANY
Amsterdam - London - New York 1973
Preface

The origin of this book is a review paper that Professor Leon Knopoff suggested be prepared for the final Upper Mantle Symposium held in Moscow in 1971. We have greatly enlarged the scope of this proposed paper.

The book is an attempt to give a broad exposition of the plate-tectonics hypothesis. We feel that, at a time when plate tectonics is often used to justify wild extrapolations from poor data with little rigor, our approach may have some value. Accepting plate tectonics as a valid working hypothesis, we try to present in a logical fashion the main underlying concepts and some related applications. The emphasis is placed on the tight constraints that the hypothesis imposes on any interpretation made within its framework. The dynamics of the plates and the origin of the motions are not discussed. There is not yet a satisfactory answer to these problems, one of the difficulties being that the rigid lithosphere is an efficient screen between us and the asthenosphere.

In its first five years of life, plate tectonics has been responsible for an extraordinarily profuse literature. It has not been our intention to provide a comprehensive bibliography of this most recent literature. We have selected about 600 references from articles and books available to us by early 1972. Most of these were chosen because we felt that they were either significant or representative of trends in research. There are no doubt some grievous omissions.

It is not possible at the present time to cover adequately the implications of the still evolving plate-tectonics hypothesis upon the different fields of earth science: Our position on many problems is controversial and partly biased. The choice of the problems is itself biased, because we have put the emphasis on those with which we, as marine geophysicists, are most familiar. After a brief introduction in Chapter 1, we define plate tectonics (Chapter 2). In Chapter 3, we describe the rheological stratification of the upper layers of the earth, defining lithosphere and asthenosphere. In Chapter 4, we discuss the kinematics of relative motions, instantaneous and finite, on a plane and on a sphere. In Chapter 5, we consider "absolute motions", that is motions within a reference frame external to the plates. Chapter 6 is concerned with processes at accreting plate boundaries and Chapter 7 with processes at consuming plate boundaries.

The book is largely the result of the close collaboration of two of the authors (X. Le Pichon and J. Francheteau). The third author (J. Bonnin) provided a first version of Chapter 7 and contributed to the general organization of the work. In many places, we have used the clear expositions of various problems that have been given by Dan McKenzie. His analytical solutions, in particular, are very convenient for discussion. We wish also to thank Jason Morgan who commented on the first four parts of the manuscript and made many valuable suggestions.
Several colleagues critically read parts of the manuscript at different stages and made constructive suggestions, particularly J. Brune, J. Cann, J. Coulomb, K. Lambeck, J.L. Le Mouel, L. Lliboutry, D.P. McKenzie, H.D. Needham, A.R. Ritsema and E. Thellier. A. Weill helped in some of the computations. K. Lambeck wrote part of Chapter 4. Many colleagues kindly gave us permission to use their illustrations and sent us papers in advance of publication. We thank Yvette Potard and Nicole Uchard for typing several versions of the manuscript with skill and style, and Daniel Carré and Serge Monti for drafting assistance. A.R. Ritsema arranged for us to delete our review paper from the Final Upper Mantle Symposium special volume and encouraged us to publish an expanded version.

This work was supported by the Centre National pour l'Exploitation des Océans, and was undertaken at the Centre Océanologique de Bretagne. We are grateful to its Director, René Chauvin, for his friendship and support. Revisions of part of this work were done by the first author while working at the Institute of Geophysics and Planetary Physics in La Jolla with a Cecil and Ida Green scholarship.

Pointe du Diable,
June 1972
Foreword


The conclusion of the active period of the Upper Mantle Project was celebrated in August 1971 by a five-days review symposium during the XVth General Assembly of the International Union of Geodesy and Geophysics in Moscow. One of the invited speakers at the time was Xavier Le Pichon, who presented a review paper on Plate Tectonics, compiled together with his co-workers Jean Bonnin and Jean Francheteau.

The broad set-up of their report made inclusion of this major review paper in the proceedings of the symposium impossible. These proceedings therefore, published in 1972 as a special issue of Tectonophysics entitled *The Upper Mantle*, and also as Volume 4 in the series Developments in Geotectonics, did not contain this paper on Plate Tectonics.

The development of the concept of Plate Tectonics and the finding of supporting evidence for the hypothesis are among the major results of the work executed during — and for a part under the auspices of — the UMP. The present Volume therefore, made up-to-date as to December 1972 with the data from literature as well as the newest results of the work of the authors themselves, may properly be considered as a direct continuation — and in effect part — of *The Upper Mantle*.

During the UMP a great deal of scientific ingenuity was directed to the quasi steady-state aspects of the earth’s crust and upper mantle. In rapid succession, new basic facts were discovered about the prominence and extent of low-velocity layers in the upper mantle and crust, and about the existence and delineation of important lateral inhomogeneities in the upper mantle. But also more dynamical aspects emerged, such as the really astonishing degree of relative motion between crustal blocks of all sizes at least during the past few 100 m.y. It is this dynamical aspect that forms the subject of the present Volume and that in large measure did lead to the initiation of the Geodynamics Project.

The problem of the driving forces for the important relative motion, and for the creation and consumption of greater and smaller plates of the earth’s lithosphere has not yet been solved conclusively. This publication therefore, is also an interim document on the present state of the art of Plate Tectonics. As such it may be expected to serve as the inspiring base for further studies in the framework of the International Geodynamics Project.

January 1973

A. Reinier Ritsema
editor of the proceedings of the final UMP symposium
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motion, as much as 1,000 years may be necessary to eliminate this statistical variation. However, the results of Davies and Brune (1971) show a fair quantitative agreement between seismic dip-slip rates obtained along consuming plate boundaries with the Brune method, using the last seventy years of data, and rates derived from those obtained at accreting plate boundaries by the Vine and Matthews method.

Over continental transform faults, the method has given rates in agreement with geodetic rates using a width $W_0$ of 20 km (about 5–6 cm/year for the San Andreas fault and 11 cm/year based on the seismicity of the last 31 years for the Anatolian fault). Over oceanic transform faults, if the motion occurs entirely by dislocation, $W_0$ is much smaller and of the order of 5 km (Brune, 1968; Northrop et al., 1970). If we can assume as a first approximation that the thickness of the brittle lithosphere increases with age away from the accreting plate boundary, as suggested by thermal models, one should make the estimation of the average $W_0$ as a function of the average age of the crust along the transform fault. For example Brune's calculations for the Romanche fracture zone actually use the lengths of three fracture zones, Romanche (880 km), Chain (330 km) and St.-Paul's (550 km). Using a spreading rate of 1.7 cm/year (Le Pichon, 1968), the mean age of the sea floor along these three fracture zones is 9.5 m.y. For the South Pacific Eltanin fracture zone complex, Herron (1972) indicates three transform faults which give a mean age of 5 m.y. using a spreading rate of 4.5 cm/year. The average thickness of 1.2 km found for the Eltanin fracture zone would thus apply to an average of 5 m.y. whereas the average thickness of 6.5 km for the Romanche applies to an average age of 9.5 m.y. Similar reasoning can be applied to the other transform faults studied by Brune (1968) and Northrop et al. (1970) and the results seem to confirm the existence of a progressive thickening of the brittle lithosphere with increasing age.

To summarize, the Brune method is, with the geodetic method on the continents, the only direct method available at present to measure relative motion at consuming plate boundaries. It is much less precise than the Vine and Matthews method. An unknown systematic error may exist if the seismicity of the period used is not representative of the “steady state” seismicity (averaged over times of the order of a million years). The main imprecision results from the uncertainty of the empirical relations giving the seismic moment as a function of surface-wave magnitude. However, this difficulty may be lessened by determining the moment directly by study of individual earthquakes, when possible, rather than by assuming a simple magnitude versus moment relationship. A second source of error is the estimation of the area of brittle faulting. The method cannot be applied to oceanic transform faults to obtain the relative velocity of plates but may be the best way to measure the variation of the thickness of the brittle lithosphere near the crest.

**Geodetic methods**

The direct measurement of motion between parts of the earth's crust is essentially

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1 This section has been written by Kurt Lambeck, GRGS, Observatoire de Meudon, Meudon, France.
one of repeatedly measuring the positions of well-defined points. As the expected motions are of the order of a few cm/year (ignoring the catastrophic displacements occurring near earthquake epicenters) the highest precision is necessary in both the measurements and in the definition of the points between which the measurements are made. Also, as we do not know if the movements are continuous and gradual, or sporadic and abrupt, we must be able to obtain the highly accurate positions within short time intervals. The elasticity of the plates requires that the measurements be made between points that are at considerable distances on either side of the plate boundary. Otherwise one is measuring the instantaneous deformation in localized areas and continuous observations over very long time periods will be required to obtain the relative motions of the plates as a whole.

Direct measurements of the positions of points on the earth’s surface can be made using various terrestrial, astronomical or spatial methods. The first ones are of limited applications, however. They require intervisibility between stations and the method is limited to measuring motions across plate boundaries located on continents. These boundaries are in general very complex and deformed over a wide region so that an elaborate network has to be constructed to ensure that at least some of the points lie on the “rigid” parts of the plates. Unfortunately, it is a characteristic of geodetic nets that the more extensive the net the less precise the results, as many circumstantial factors become important. Spatial geodetic methods circumvent this problem because one can measure the positions of points separated by several thousands of kilometers. That is, the points can be selected to lie far from the plate boundaries and motions across oceanic plate boundaries can be measured. The spatial methods have only been developed in recent years and they do not yet provide the high accuracy necessary for measuring the tectonic movements. Nevertheless they are very promising and will probably give the required results in the near future. Terrestrial and spatial geodetic measurements are complementary; the former can provide the motions occurring immediately across the boundaries and the latter can provide the motions of the plate as a whole. These two types of displacements are required for a complete understanding of the tectonic motions.

In satellite geodesy one often speaks of relative and absolute position determination. Positions are considered “absolute” in that they refer to an inertial reference frame having for origin the earth’s center of mass. This inertial frame can be defined by a z-axis parallel to the earth’s mean axis of rotation, an x-axis in the plane of the mean equator along the vernal equinox at an adopted epoch and a y-axis along z x x. Given points on the earth’s surface can be related to this frame if the motion of a frame attached to the earth’s crust about this inertial frame is known. That is, if precession, nutation, polar motion and variations in the earth’s rate of rotation are known. The motion of an earth satellite is described with respect to this inertial frame and, with the dynamic method of satellite geodesy for example (see below), satellite tracking stations are determined in the inertial system. These station positions are of course time-dependent due to the earth’s motion and they can be related to some
terrestrial frame only if the above mentioned motions of the earth are known. Any variation in the position of the station due to tectonic motion manifests itself as a change in the coordinates of the point with respect to the terrestrial frame. The difficulty is of course two-fold: the motions of the earth are not known with sufficient accuracy and will have to be improved with the same spatial techniques as used for measuring the tectonic motions, and the terrestrial frame can only be pragmatically defined by the coordinates of the stations themselves. A small displacement in one station will change the definition of the terrestrial frame and any concept of absolute positions becomes meaningless at the level of accuracy sought here.

The spatial methods that are likely to be of interest in the near future are the precise tracking of close-earth satellites, laser-range measurements to the moon and radio-telescope long-baseline interferometry. In view of the above discussion these methods will give relative motions only. They are in many cases complementary and will probably have to be considered together in any future observing campaign.

If the distances to satellites are measured simultaneously from several stations it is possible to determine the relative station positions by a purely geometric method, without recourse to any orbital theory. This method is usually referred to as the geometric method of satellite geodesy and has been successfully used in the past for simultaneous direction measurements and for simultaneous direction and distance measurements. The accuracies obtained have generally been of the same order as the accuracy of the observational data; about 5–10 m for direction measurements only (Gaposchkin and Lambeck, 1971) and about 2–5 m for the direction and distance measurements (Lambeck, 1968; Cazenave and Dargnies, 1971). Relative positions of points separated by several thousand kilometers can be determined in this way.

For improvements beyond this level of accuracy only the simultaneous observations of laser distances will provide the means. The method has not yet been fully used because many stations are required to give precise and unambiguous results. For measuring the motions of some of the large plates, as many as six stations, three on each plate, are required to give the complete motion and to ensure that any relative motions between the points on the plate can also be detected. It is not, of course, necessary that all six stations observe the satellite at the same time. Simultaneous observations from subgroups of any four stations at a time will provide important information. For measuring the motion of small plates relative to big plates, four stations will suffice, three on the main plate and the fourth on the small plate, provided there is no motion between the stations on the principal plate. With the geometric method being based on very simple hypotheses, the station positions can be determined with an accuracy comparable to that of the observations particularly as the station–satellite configuration can be optimized.

Station positions can also be determined by the so-called dynamic method of satellite geodesy. The forces acting on the satellite are assumed known or partially known so that the satellite’s positions can be computed at any instant of observation as a function of known and unknown parameters. Because the equations of the satellite’s
motion are referred to a well-defined inertial frame centered on the earth’s center of mass, these satellite positions also refer to this inertial frame. The observations of the satellite positions and the computed geocentric positions are related to the unknown or partially known geocentric station positions, giving a set of equations whose solution yields, amongst other information, the correction to the station’s position. The difficulty with this method is that we need to know all the forces acting on the satellite and that we have to know very precisely the motion of a frame attached to the earth’s crust relative to the inertial frame. At present, an accuracy of between 5 and 10 m has been obtained using observational data of about 10–15 m accuracy (Gaposchkin and Lambeck, 1971). The advantage of the dynamic approach is that the satellite does not have to be simultaneously visible from several stations at a time. This means that the separation between stations can be very much greater than for the geometric approach and that fewer stations are required to completely measure the motions of the major plates.

The immediate goal in satellite geodesy is for observational accuracies of about 20 cm in station–satellite distance and for this we can use the existing satellites, particularly GEOS 1 and GEOS 2. Accuracies of about 20–50 cm in station position could be achieved within about three years from now (1972) if a suitable observation campaign is organized. These accuracies are still not very interesting for measuring tectonic motions but they are most valuable for other interactions between satellite geodesy and geodynamics. Beyond this level of accuracy we need to know the earth’s motion relative to an inertial frame with accuracies that are higher than the presently used methods can provide. To achieve accuracies better than about 10 or 15 cm, special satellites will be required to ensure that we can establish the point to which we make the measurement and to minimize, for the dynamic method, the non-gravitational forces. Such a satellite has been proposed by Weiffenbach (1970) and is now being studied in detail (Weiffenbach and Hoffman, 1970) for a possible launch date in 1974. The altitude of the orbit will be between 3,000 and 4,000 km and the optimum distance between laser stations for the geometric solution is of this order.

**Lunar laser ranging.** Distance measurements between laser stations on the earth and retroreflectors on the moon are of the same order of accuracy as achieved by ranging to close-earth satellites; an accuracy of about 10 cm is envisaged in the near future. Instrumental complexity is, however, very considerably increased, as the moon is so much further away. For example, the McDonald Observatory group (Alley et al., 1970) is using a 2.7 m telescope for transmitting and receiving the laser beam, whereas for close-earth satellite tracking a 40 cm transmitting optics is quite adequate (Lehr and Pearlman, 1970).

The reflectors on the moon could be used in a geometric manner exactly as discussed above for close-earth satellites. However, the geometry is considerably poorer now, and reliable results can only be obtained if the terrestrial stations are well distributed around the entire globe. Bender et al. (1968), for example, have investi-
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gated the geometry for very high satellites (maximum distance 110,000 km) and conclude that with 12 globally distributed stations and with range accuracies of about 15 cm, it is possible to measure interstation distances with accuracies of about 40 cm. For the moon these results will be poorer. An alternative method, similar to the dynamic method of satellite geodesy, is to use the moon’s orbit as reference for a short time period. Alley and Bender (1968) discuss this approach for measuring relative longitudes between two stations. The method is to observe distances to the reflector at times about equally distributed about the moment when the moon passes through the local meridian. The difference of these measurements gives a measure of the time at which this passage occurs. Repeating the measurements from a second station gives the difference in longitudes of the two stations. We need to know, however, the precise variations in the motion of the reflector on the moon for the period of observation (usually about 8 hours) to obtain a good geometry. It is not at all clear with what accuracy we will be able to predict this motion in the future. We need also to know the polar motion, the variations in the earth’s rate of rotation and the earth and moon tides. Separating these various phenomena will again require many well distributed stations. It would seem that the lunar laser ranging methods are most suitable for studying the motions of the moon and perhaps the earth’s rotation. For tectonic motions, the use of laser ranging to artificial satellites would be more appropriate.

One of the limiting factors in laser ranging to objects outside of the earth’s atmosphere is the retardation of the laser pulse by the atmosphere. The effect can be corrected for, but uncertainties of a few cm may remain. Studies by Hopfield (1972), however, are very encouraging in this respect: comparisons of refraction corrections computed from model atmospheres and surface measurements with correction based on atmospheric soundings indicate agreement to better than 1 cm near the zenith.

*Long-baseline radio interferometry* (V.L.B.I.). In the classical interferometric methods a radio signal is received simultaneously at two antennas separated by a known baseline. Because of the different path lengths between these two terminals and the radio-signal source, the two signals received will exhibit a phase difference that is a measure of the path length difference of these signals. If the baseline is known the orientation of the source relative to the baseline can be determined.

To measure the phase difference the received signals have to be compared against a standard oscillator. In the past, when the local oscillators were not very stable, the two antennas were connected electrically and the signals compared to the same frequency standard. This need for a direct connection limited very much the maximum length of baseline that could be achieved because of electronic delays in the land line. This need also meant a limit to the resolution that can be obtained since the longer the baseline the larger the phase difference and the more precise the determination. Since the development of stable atomic clocks these problems can now be overcome. The received signal at each terminal is now recorded on tape, together with the time control derived from the frequency standard. At some later date, the two tapes can be brought...
together and analyzed for the phase difference. Now the length of the baseline is only limited by the fact that the source must be simultaneously visible from the two antennas. The method has been most successfully applied in resolving stellar sources; that is in measuring difference in angles. Cohen et al. (1968), in a summary of results, report resolutions of better than 0.001 sec of arc for a baseline of 6,300 km. The inverse of those results would be that we can obtain the baseline with a comparable accuracy if the source positions are known and if we can overcome several other difficulties.

Generally we will not know the source positions to solve for baseline and source positions at the same time. Also, we have to improve the stability of the clocks. For the source resolution measurements, it is only necessary that the local oscillators are stable for the period of observation and drifts from one period to the next are unimportant as we are making relative measurements. For the baseline determination, we require much more stringent phase control. Even hydrogen-maser frequency standards may not be adequate for centimeter accuracies. Atmospheric refraction also is a problem, more serious than with laser measurements, because at radio frequencies both the tropospheric water vapor content and electron densities in the ionosphere become important. If present available atmospheric models are used, the uncertainty in the refraction correction is of the order of 10–20 cm and is dependent on the frequency of the radio-source. Improvements can be expected if the atmospheric densities can be measured at the time of observation (Dickinson et al., 1970).

Up to now the results for baseline determination have provided accuracies that are similar to those obtained by satellite geodesy methods. Recent results by Cohen and Shaffer (1971) between stations in Australia and on the west coast of the United States compare very favorably with results obtained by Gaposchkin and Lambeck (1971) for the coordinates of nearby satellite tracking stations that have been connected to the corresponding radio-telescopes by ground survey.

So far we have not said anything about the nature of the radio-sources. Several possibilities exist. Stellar sources are most convenient because they are already there and because many of them lie outside the galaxy and are not expected to exhibit measurable proper motions (relative motions of the stars). Thus they form an ideal inertial reference frame with respect to which the earth's motion in space can be measured. The disadvantage of natural sources is that the signals received on earth are quite weak requiring very large telescopes in order to observe them. This means not only that there may be some difficulty in relating the electronic center of the instrument to well-defined points on the "rigid" earth with better than centimeter accuracy, but it also means that the use of V.L.B.T technique will be a costly way of measuring tectonic motions; particularly as several stations will be required on each plate in order to separate out the various other geophysical phenomena introducing variations in station positions. Artificial radio-sources can also be used. Michelini and Grossi (1972) for example have attempted to use radio-signals from the geostationary satellite ATS-5. Radio sources have been placed on the moon by some of the Apollo missions.
The practical advantage of these approaches is that the signals received are much stronger so that very much smaller radio-telescopes can be used and that the analysis of data is simpler and less costly than it is for stellar source observation. On the other hand, we will have to know precisely the forces acting on the satellite.

Terrestrial geodetic measurements. A simple technique of measuring displacements along fault zones is to establish a line of survey marks across the fault and to measure the offsets of the line. This can readily be done with an accuracy of a few millimeters so that it offers a very simple and accurate method of measuring fault-creep slippage along the San Andreas fault and associated faults (Tocher, 1960; Nason and Tocher, 1970). The latter, for example, report motions of about 1 mm/month across the Calaveras fault, a fault associated with the San Andreas system. Continuous recording of the movements is possible, and measurements are now being made at some 20 points along various sections of the Californian fault system. However, fault zones are not always confined to very narrow zones and we will often require more sophisticated methods.

Classical triangulation methods, using theodolite direction and invar tape distance measurements, have been used in the past to detect tectonic motions along fault zones. An area much investigated in this manner is the San Andreas fault (Whitten, 1960, 1970; Meade, 1966). Meade, for example, gives results for a survey made in 1944 and resurveyed in 1966. The displacements along the fault are quite convincingly established and in good agreement with fault-creep slippage rates known to occur on the fault; the difference in positions for points on either side of the fault amounted to 60 cm in 20 years. These methods give convincing results only when the time interval is about 20–30 years. More precise methods are required so that the displacements can be determined over shorter time intervals, in order to obtain a clearer idea as to whether these motions are continuous or sporadic. Electronic distance measurements provide a means of doing this. Conventional geodimeters, for example, give accuracies of about 2–3 parts in $10^6$, or about 4–6 cm in distances of about 20 km. A geodimeter traverse criss-crossing the San Andreas fault system for more than 600 km has been established by the California Department of Water Resources (Hoffmann, 1968). Parts of the traverse have been remeasured several times during the last 10 years and a general movement of about 2–4 cm/year has been found along the fault although many anomalous displacements occur.

Further improvements are possible with laser instruments and many tests have indicated that a precision approaching 1 in $10^7$ is possible, but that, as for the spatial measurements, the atmospheric refraction limits the accuracy to below this value. The problem is not insurmountable if the atmospheric density along the ray path can be observed. But the problem is not insurmountable if the atmospheric density along the ray path can be measured, either by discrete sampling or by measuring the integrated density along the
Astronomical observations of latitude. Observations of astronomic latitude can, in principle, also provide a means of testing the plate-tectonics hypothesis. The astronomic latitude of a station is defined as the complement of the angle between the earth's instantaneous axis of rotation and the station's vertical, and is observed, for example, by measuring the zenith distance of a known star as it passes through the station's meridian. Variations in the astronomic latitude are therefore caused by polar motion, by earth tides and by tectonic motions of the stations. The difficulty lies in separating these various components. The polar motion has periodic parts as well as a possible secular drift, and the former can be eliminated by suitably averaging the latitude data. The earth tide effects can usually be adequately modelled, as far as the short-period terms are concerned, but this cannot be said for the long periods, such as the 18.6-year period, as the earth's elastic response to long-period deformations is poorly understood.

Latitude observations have been made for some 70 years by five stations of the International Latitude Service but the results are quite inhomogeneous because of variations in methods of reduction and observation. The latitude data at each station are averaged over a 6-year interval to eliminate the annual and 14-month Chandler period. The interpretation of the variations in the mean values for each of the stations depends on the hypothesis made. Markowitz (1968), for example, analyzed the data assuming that the stations are fixed with respect to each other and found a secular drift of about 0.003°/year along longitude 65°W. Whitten (1970) on the other hand, interpreted Markowitz's results assuming there to be no secular motion of the pole. He found that it was possible to explain the variations in mean pole position if North America had turned about 5° clockwise and Eurasia about 5° anticlockwise during the last 107 years. These results are almost identical to those found by Le Pichon (1968) from an altogether different method.

The accuracy for a mean latitude averaged over a year is about 0.02" (or 0.6 m) according to Markowitz (1968). Thus at least 50 years of good data is required to observe any drift between two stations on different plates. For a separation of the tectonic motions from the secular motion of the pole, many more stations are required than now participate in the International Latitude Service observing program.

Length of sinking zone method

Isacks et al. (1968) demonstrated that the seismic activity within deep and intermediate seismic zones shows a well-defined relative maximum in some depth range in the upper mantle. They suggested that the length of the seismic zone, from the surface to its maximum in activity, is a measure of the amount of underthrusting during the past
10 m.y. The suggestion was based on the observation that the time which has been necessary to thrust into the mantle a length of plate equal to the present length of the seismic zone, using the calculated slip rates of Le Pichon (1968), did not vary with the calculated slip rate but remained close to 10 m.y. (Fig.19).

Fig.19. Calculated rates of underthrusting (Le Pichon, 1968) versus length of seismic zones. Crosses indicate unusual deep events. Note the approximately linear increase in the length of the zone with respect to the calculated slip rate. (After Isacks et al., 1968.)

The most remarkable correlation between length of seismic zone and predicted slip rate exists for plate boundaries along which the distance to the pole of relative motion varies greatly. An example strikingly illustrated by McKenzie is the Pacific–India plate boundary from Macquarie Island to Tonga where the length of the seismic zone increases from less than 60 to 800 km as the predicted rate increases from 0 to 8 cm/year (McKenzie, 1969a; see Fig.73). Another example is the Pacific–Philippine plate boundary from the Palau–Yap trenches to the Izu–Bonin trench (Katsumata and Sykes, 1969).

Isacks et al. suggested two possible explanations for this correlation. Either the present seismic zones were created during the most recent 10 m.y. old episode of spreading, or 10 m.y. was the thermal time constant of the plate within the seismic zone. The first explanation can now be ruled out, as the JOIDES results (Maxwell et al., 1970) have demonstrated that there was no worldwide pause of spreading during the last 70 m.y. McKenzie (1969a) has shown that the second explanation is possible.
If one assumes that there is a temperature limit \( T \) beyond which the material does not behave anymore as a brittle solid, the plate will lose its identity within the mantle once it has reached this temperature. The time necessary for the plate to reach \( T \) is independent of the consuming rate, provided that the mantle is at a constant temperature and that the change in pressure of the plate as it sinks does not produce a change in temperature (McKenzie, 1969a). These assumptions are not valid due in particular to pressure-induced phase changes and adiabatic heating, but they lead to a reasonable approximation if one reasons in terms of potential temperatures. In addition, the thickness of the plate as it enters the consuming zone may be quite variable depending on its age and the time necessary to heat it will vary roughly as the square of its thickness (W.J. Morgan, personal communication, 1972). In conclusion, this explanation would assume that the time necessary for a plate to reach the temperature \( T \) is about 10 m.y.

The difficulties in using this method are numerous. First, it is based on an empirical correlation which can be explained, but not yet predicted, by the progressive heating of the plate as it sinks. Too many variable parameters are poorly known or unknown: actual geometry of plate, thermal constants, pressure-induced phase changes, distribution of temperature within the mantle. Second, one has to assume that the plate did not begin to be thrust into the mantle later than 10 m.y. ago. This is not necessarily true. Third, the length of the zone should be measured along the direction of slip, which should be estimated from fault plane solutions. Fourth, Fig.19 shows that the empirical relationship between rate of slip and length of zone is poorly defined. This method should, in fact, be used to obtain a better knowledge of the thermal time constant, and consequently the thickness, of the different deep seismic zones. Such a study is possible now that we have much more detailed data on the geometry of the deep seismic zones (Isacks and Molnar, 1971) and better calculated rates of slip (Morgan, 1971b). In this study, one should take the age of the lithosphere into account. It is already very suggestive that the Japan seismic zone, in which the lithosphere is very old (> 200 m.y.), is the longest and the Middle America seismic zone, in which the lithosphere is very young (a few million years), is among the shortest (Molnar and Sykes, 1969). Clearly, this study would set important needed constraints on the thermal models of the sinking plate.

**Methods of measurement of direction of relative motion**

**The transform-fault method**

Transform faults are plate boundaries along which surface is conserved and consequently lie along the direction of relative motion between plates. A knowledge of their geometry uniquely defines the direction of motion. This method is very simple to apply and quite powerful. Heezen and Tharp (1965) were probably the first who suggested that the motions of the continents on either side of a mid-ocean ridge may be inferred from the direction of fracture zones. Morgan (1968) and Le Pichon (1968)
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used transform faults to obtain the directions of relative motion between plates in a quantitative way.

Ridge-ridge and ridge-arc transform faults have the property that a residual inactive trace exists in their prolongation, beyond the accreting plate margin. Thus the active part gives the present direction of relative motion while the fossil part gives a geological record of the past relative motions of the two plates through time. This type of transform fault generally lies under water in the ocean and has a very clear topographic expression. Menard (1955) first described these topographic features in the Pacific Ocean and called them fracture zones, the name being applied to both the active and fossil parts of transform faults. Menard and Chase (1970) define them as “long and narrow bands of grossly irregular topography characterized by volcanoes, linear ridges and scarps, and typically separating distinctive topographic provinces with different regional depths”. Fracture zones have now been discovered along the whole length of the accreting plate boundaries. Their length, their size, and the offset of the ridge crest along them are highly variable. In general, the size of the topographic feature increases with the length of the offset and this is understandable on mechanical and thermal considerations. The length of the fracture zone shows some correlation with the length of the offset. This results from the fact that transform faults with large offsets are very stable, whereas transform faults with small offsets are not and may disappear due to second-order modification of the geometry of the accreting plate boundary. However, no clear correlation has been found between relative velocity along the boundary and size of the fracture zone.

Typically, fracture zones consist of a linear trough a few km (~10 km) wide and a few hundred meters deep. This trough is often obscured or filled by sediments and may be difficult to map without a seismic reflection survey. The trough may be bordered by one or two basement ridges or scarps, which may have a height of a few kilometers and are easy to map. The total width of a large fracture zone is about 30 km. However, the structure is often complex and many exceptions to the preceding description exist, as shown by Fig.20. The magnetic field is generally featureless over the fracture zone and this has been explained by Cann and Vine (1966) by a demagnetization of the rocks due to shear metamorphism. On each side of the zone of quiet magnetic field, the Vine-Matthews magnetic lineations are offset by a distance equal to the offset of the ridge crest at the time they were created. This provides probably the surest way to identify and map a fracture zone, especially if the offset is not large. The precise locations of epicenters can also be used to determine the geometry of the active part of a fracture zone (Sykes, 1963, 1965, 1966a, 1967; Sykes and Landisman, 1964).

The major difficulty in surveying a fracture zone is that it is possible to go from one fracture zone to a parallel one, from one crossing to the other. This is especially true if there are numerous closely spaced fracture zones of equal importance and where they are disrupted by an important change in trend. Once the fracture zone is mapped, it is necessary to determine the exact line of slip. Yet, it is not clear which part of the
The fracture zone can be taken as the exact witness of the path of a flow line and how much the fracture zone can deviate locally from a flow line. Francheteau and Le Pichon (unpublished) have shown that, in the Gibbs (or Charlie) fracture zone, near $52^\circ$N in the Atlantic Ocean, the axis of the trough follows the same small circle to within 3 km over long distances, whereas the scarps on each side may deviate from a small circle by as much as 10 km locally (Fig. 21). In general, the fracture zones define flow lines to within 3–5 km. The effects of the small deviations from latitude circles are probably absorbed by the elasticity of the plates. These deviations, however, put a limit to the resolution of the direction of relative motion one can get from a fracture zone.

Theoretically, the informations contained in the geometry of the flow line followed by a transform fault is sufficient to determine uniquely the location of the pole of relative motion. This results from the fact that the curvature of a small circle varies with the distance $\theta$ to the pole of rotation as $1/R \sin \theta$ where $R$ is the radius of the spherical earth. Thus, the variation of the curvature obeys the same law as the variation of the rate of accretion and contains the same information as far as the location of the pole is concerned. In practice, it is very difficult to define precisely the curvature of the latitude circle along the small portion of an active transform fault, unless one is very close to the pole of rotation. Consequently, one is generally only able to define a locus of the pole, which is the great circle perpendicular to the portion of the active transform fault. The only useful information then is the azimuth of the transform fault, which can be considered as the derivative of the different mapped positions of the fracture zone. Locally, this azimuth may be greatly in error and a smoothing method is necessary. Many authors (Morgan, 1968; Le Pichon, 1968) have used a visual smoothing method. It is probably best to do the smoothing numerically over the different positions taken at equal interval along the corresponding portion of a transform fault. This will be discussed more completely later in this chapter (p. 115).

If transform faults associated with accreting plate boundaries are easily identified, this is not so in the case of consuming plate boundaries, where it is not possible to recognize whether a topographic trench is due to thrusting of one plate below the
Fig. 21. Topography of the Gibbs (or Charlie) fracture zone and computed small circle adjusted to the points shown by black dots. (After Franchette and Le Pichon, unpublished work.)
other or to pure strike-slip. Thus, along consuming plate boundaries, earthquake fault-plane solutions and geodetic measurements are the only possible methods.

**Fault-plane solution method**

Along a plate boundary, the movement on the fault plane is mostly the result of successive dislocations which occur each time the accumulated strain exceeds the plate's elastic limit. The opposite sides of the fault then return to a position of equilibrium, releasing in elastic waves a great part of the accumulated energy. For an earthquake of magnitude larger than 5.5–6, these waves can be recorded over the whole earth. The first motion of a compressional P-wave can be either a compression or a rarefaction. The distribution of these first motions, as recorded at different stations surrounding the epicenter, can be used to obtain the sense and type of displacement which occurred on the fault plane. The “fault plane” or “focal mechanism” solution is a powerful way to determine the direction of relative movement between plates, in spite of the fact that the precision of a given solution rarely exceeds 10°–15° and that there is an inherent ambiguity of 90° which must be raised on the basis of other considerations. The technique was perfected in the 1960’s (Stauder, 1962; Wickens and Hodgson, 1967) and the results showed that all well recorded earthquakes could be satisfactorily explained by the equivalent double-couple point source (Honda, 1962). The technique was first applied to plate tectonics by Sykes (1967) who demonstrated its remarkable efficiency with the use of the long-period seismograph records of the World Wide Standardized Seismograph Network which commenced operation in 1962. Previously, solutions showed a large proportion of inconsistent readings, due to the non-homogeneity of the network. In addition to the P-wave onset method, fault-plane solutions can be obtained from the direction and polarization of S-wave onset (e.g., Stauder, 1962), from surface waves (Brune, 1961) and from the amplitude of free oscillations data (Gilbert and McDonald, 1961). These last two methods are more difficult to use and less reliable than the methods based on onsets of P- and S-waves. In the following, we will briefly discuss the P-wave onset method, making extensive use of a very clear discussion by McKenzie (1971b).

Let us consider a purely horizontal motion on a vertical plane, the sense of motion being indicated by the arrows on Fig. 22. Intuition suggests that points in front of the arrows are being pushed while points behind them are being pulled. This is confirmed by dislocation theory. In directions such as B and E, the initial motion is away from the focus of the earthquake (compression) while in directions such as A and F it is toward the source (rarefaction or dilatation). The radiation field is divided by two perpendicular planes into dilatational and compressional quadrants. These planes, called nodal planes, are planes along which in theory the motion of P is null. This property, actually, helps to determine whether the station where the waves are recorded is close to a nodal plane. One of the nodal planes is the fault plane, the other is the auxiliary plane which is perpendicular to the slip vector. It is not possible to decide from the wave-radiation pattern only which plane is the fault-plane. This ambiguity is funda-
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Fig. 22. The radiation field of a strike-slip earthquake. The arrows on the rays mark the initial direction of motion of the ground. Note that the auxiliary plane is perpendicular to the slip plane. (After McKenzie, 1971b.)

mental to the double-couple source mechanism. The decision must be made on the basis of other informations: the earthquake may have been accompanied by a surface break and the strike of the break as well as the sense of motion along it must agree with one of the two possible solutions. The distribution of aftershocks, the ellipticity of isoseismal lines and the direction of propagation of the initial break are other possible ways of removing this ambiguity.

Let us define the normal to the auxiliary plane (called its pole) by the unit vector \( u_1 \), the normal to the fault plane by \( u_2 \) and the unit vector along the vertical by \( a_z \). The slip vector in the fault plane is parallel to the pole of the auxiliary plane \( u_1 \). The horizontal projection of the slip vector then is \( u_h = u_1 - (u_1 \cdot a_z) a_z \) and consequently, the strike of \( u_h \) may be easily obtained by adding 90° to the strike of the auxiliary plane. If the auxiliary plane is vertical, as in Fig. 22, the strike of \( u_h \) is the same as the strike of the fault plane. But this result is not true otherwise and the strikes of \( u_h \) and of the fault plane in the horizontal plane will in general be different. McKenzie and Parker (1967) showed that, along a given plate boundary, one can use the horizontal projection of the slip vector to raise the ambiguity between fault plane and auxiliary plane. This results from the fact that \( u_h \) should have a consistent direction as one moves along the plate boundary: the rigidity of plates imposes continuity in direction of relative movement but the fault plane itself may have any direction. Thus a choice between the two planes may be made on the basis of the necessary continuity of \( u_h \) along the plate boundary. The intersection of the two planes, called the null vector or \( B \)-axis is in the direction \( u_1 \times u_2 \). The \( P \) (or Pressure) axis lies in the dilatational quadrant, is normal to the null vector and bisects the two planes while the \( T \) (or Tension) axis similarly bisects the compressional quadrant and is normal to the null vector; written as unit vectors they lie along \( (u_1 + u_2) / \sqrt{2} \) and \( (u_1 - u_2) / \sqrt{2} \). Within an homogeneous material the triaxial stress-field applied to the material should correspond to \( P, B \) and \( T \), where \( P \) is the axis of maximum compressive stress, \( B \) the axis of intermediate and \( T \) of minimum compressive stress. It has been shown by Isacks and Molnar (1971) that this is true within the plates sinking in the mantle in which earthquakes occur by failure within an homogeneous material. However,
McKenzie (1969b) has pointed out that this is not true of earthquakes produced by differential motion between plates. This is because the fault plane is already a pre-existing plane of weakness and that shear stresses involved in shallow earthquakes are at least an order of magnitude too small to produce fracture within an homogeneous fault-free material. Thus, in these cases, failure may occur at an angle different of 45° from the axis of greatest stress and the $P, B, T$-axes have no simple relation to the triaxial stress field.

In practice, to obtain a fault-plane solution on the basis of records of the radiation wave pattern at stations far away from the earthquake, it is necessary to know the directions in which the rays left the focus. But the path of the ray depends on the velocity structure of the earth along its path, on the angular distance between the source and the receiver and on the hypocentral depth. Thus, the calculated angle $i$ between the rays and the downward vertical may be affected by sizeable errors, especially for crustal earthquakes which occur in regions of large velocity gradients and if the structure is not radially homogeneous. This will be specially true if the nodal planes are not steeply dipping as in a strike-slip solution. Values of $i$ for a given distance and hypocentral depth can be obtained from tables (see Ritsema, 1958; Sykes, 1967). Allowance for a velocity in the focal region less than 7.8 km/sec (crustal

**Fig. 23.** Example of a nearly pure strike-slip mechanism for an event on a North Atlantic fracture zone (event 5 of Sykes, 1968). Solid circles: compressions; open circles: dilatations; crosses: wave character on seismogram indicates station is near nodal plane. Smaller symbols represent poorer data. $\phi$ and $\delta$ are strike and dip of the nodal planes. Arrows indicate sense of shear displacement on the plane that was chosen as the fault plane. (After Sykes, 1968.)
earthquakes) is made very seldom, although it could easily be done, using for example Ritsema's curves.

Knowing the angle $i$, a convenient way to represent the radiation pattern in two dimensions is to imagine a small sphere centered on the focus of the earthquake on which the first-motion directions are plotted and to project the lower hemisphere onto a horizontal plane using a stereographic or an equal-area projection. In the latter case for example, the two coordinates used to describe a station are its azimuth at the epicenter and its radial distance $R$ where $R = \sqrt{2} \sin (i/2)$, using a sphere of radius unity. Thus, the intersection of the sphere with the plane of projection is the circle $R = 1$. Fig.23–25 show three examples of representations of fault-plane solutions respectively for dominantly strike-slip, normal faulting and thrust-faulting solutions. It is easy to see that the strike of a nodal plane in the horizontal plane is obtained by measuring the angle on the circle of radius $R = 1$ clockwise from the north. The strike of $u_h$ is obtained by adding 90° to the strike of the auxiliary plane. The complement $i$ of the dip of a plane is obtained by measuring the distance $R = \sqrt{2} \sin (i/2)$. For a "pure" strike-slip solution with a vertical fault plane, the circle is divided into four

![Fig.24. Example of a nearly pure normal-faulting solution of an East African earthquake (event 13 of Sykes, 1968). Note that the two nodal planes have approximately the same strike and that consequently, the horizontal projection of the slip vector is uniquely defined (with a possible error of 18°); $C$ and $T$ correspond to the $P$ and $T$-axes. $R = \sqrt{2} \sin [(90°-6)/2]$. (After Sykes, 1968.)](image-url)
Fig. 25. Example of a nearly pure thrust-faulting solution of an earthquake on Macquarie Ridge (event 17 of Sykes, 1968). (After Sykes, 1968.)

equal quadrants and the motion along a nodal plane is such that it goes from a dilatational to a compressional quadrant. For a purely normal-faulting solution, the center of the circle is occupied by an oval filled with a dilatational quadrant. For pure thrust-faulting the oval is filled with a compressional quadrant. For these last two cases, there is no ambiguity in determining the strike of \( u_h \) as the two nodal planes have the same strike. However, a combination of strike-slip with normal-faulting and thrust-faulting is possible.

This method is the most general method yet devised to study motions between plates as it applies to any plate boundaries. Since Sykes (1967), it has been widely applied by many authors and first by McKenzie and Parker (1967) along a consuming plate boundary and by Isacks et al. (1958) over the whole earth. The precision obtained with it is routinely of the order of 20° for an earthquake of magnitude 6 or greater, and could normally be improved considerably if a proper distribution of recording stations existed and if the upper structure at the epicenter were well known. However, the distribution of recording stations on the focal sphere is most often very inhomogeneous (Davies and McKenzie, 1969).

There is an interesting application of this method to micro-earthquake studies using
local recording networks. It would be of great interest to compare the results obtained to those corresponding to large earthquakes.

**Calculation of instantaneous relative angular velocity between plates and estimation of errors**

*Introduction*

The instantaneous relative angular velocity between two plates can be represented by a pseudo-vector \( \mathbf{\omega} \). The problem of determining the relative motion between two plates on the earth’s surface then reduces to the calculation of three parameters, the coordinates of \( \mathbf{\omega} \), and their probable errors. Alternatively, the determination of \( \mathbf{\omega} \) can be obtained in two steps. One first computes \( k \), the unit vector along \( \mathbf{\omega} \) (entirely determined by two parameters, i.e., the ratios of two cartesian coordinates with respect to the third one). This is equivalent to finding the location at which the rotation axis pierces the earth’s surface, which is defined by two parameters (latitude \( \phi \) and longitude \( \lambda \)). Then one has to obtain the third parameter \( \omega \), the magnitude of \( \mathbf{\omega} \).

In practice, to make this determination, we have a population \( I \) of points along the common boundary of the two plates at which the magnitude \( v \) and (or) the direction \( v/\nu \) of the relative velocity vector \( v \) at a point \( r \) have been measured. We will define \( u = v/\nu \). It is important to note that these two sets of measurements, \( v \) and \( u \), are independent and are not obtained in general at the same points. Thus there are two sets of errors, those on \( u \) and \( u \) and not simply one, the errors on \( v \). In addition, these errors are often difficult to appreciate and systematic errors may be present. For example, fault-plane solutions are often affected by unknown asymmetry around the earthquake focus, especially if the fault plane is not vertical. Finally, as the data are generally discontinuous and inhomogeneous, it is not easy to test their internal consistency.

Very little systematic work has been done up to now on the problem of determining \( \mathbf{\omega} \) from the \( v \) and \( u \) measured for \( I \) and evaluating its probable error. McKenzie and Parker (1967) apparently determined the America/Pacific unit rotation vector \( k \) by construction of great circles perpendicular to the horizontal projections of slip vectors on a globe. Morgan (1968) obtained \( \mathbf{\omega} \) from a bootstrap operation combining a similar but computerized geometrical construction with considerations on distances to poles based on the law of variation of \( v \). He first obtained the location of the eulerian pole \((\lambda, \phi)\) by geometrical construction of great circles perpendicular to \( u \) at point \( r \) (Fig. 9, upper part). The great circles tend to bundle in an area which is elongated along the general direction of the great circles. Then a visual inspection of the variation of \( v \) with distance to the chosen pole (Fig. 9, lower part) led him to choose a “best” position for the eulerian pole within this elongated area. This then fixed the value of both \( k \) and \( \omega \).

Le Pichon (1968) used instead two computerized search methods. In the first method, the \( k \) chosen was such that the function \( F = \Sigma(\theta - \psi)^2 \) be minimized, where \( \theta \)