Horizontal Lithosphere Compression and Subduction: Constraints Provided by Physical Modeling

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Physical modeling of the subduction is performed with a two-layered mantle model (elastico-plastic lithosphere and low-viscosity asthenosphere) and is governed by the criteria of similarity. Compression of the lithosphere in the area of a passive continental margin has been shown to produce a buckling instability in the oceanic plate with wavelengths of 200 km on the average. Later, a localization of deformation occurs in sagging at some distance from the margin where a strongly dislocated linear ridge is formed due to the thrusting. The plate then experiences a failure along the inclined zone, and subduction starts. The inner trench slope which forms therefore has a scraped structure and include a block of crushed and dislocated oceanic crust and sediments in the lower section. If there is an old inclined fault striking across the compression of an oceanic lithosphere, it is on that fault that a subduction zone is initiated. The inner trench slope has then a different structure and forms, due to normal faulting, in the frontal part of the overriding plate. The formation of a subduction zone requires a compression that is smaller than that in the preceding case by a factor of 2 or 3. The subducting plate experiences an elastico-plastic bending and (under specific conditions) thrusting along the zone, dipping from under the overriding plate oceanward and crossing the entire lithosphere. The best agreement between generalized relief in subduction zone in the model and nature is achieved when a shear yield limit \( \sigma_s = 1.3 \times 10^8 \) Pa, modulus of elasticity \( E \) about a few times \( 10^{11} \) Pa, and a thickness \( H = 60 \) km, are adopted for the real lithosphere.

I have also done some joint work with colleagues in the physical modeling of subduction and accompanying processes [Shemenda, 1979, 1981, 1984, 1985, 1989a,b,c; Grocholsky and Shemenda, 1985; Lobkovsky and Shemenda, 1981; Lobkovsky et al., 1980] using specially fabricated hydrocarbon compositional systems to model the lithosphere. Modeling was performed with a two-layered mantle model involving elastico-plastic lithosphere and low-viscosity asthenosphere. A similarity criteria have been satisfied in the modeling. The present paper is a further extension of this work. It contains a description of the conditions and results of mechanical subduction modeling in an environment of horizontal compression.

GENERAL MODELING SCHEME

The following factors should be taken into consideration when determining the general scheme of mechanical modeling of subduction. The most important thing is the fundamental feature of upper mantle structure, namely, its two layers. The upper layer, the lithosphere, consists of stronger (or more viscous) material, while the lower layer, the asthenosphere, has a strength (viscosity) that is several orders lower. Second, the lithosphere in the transition region from ocean to continent where subduction zones often develop is strongly heterogeneous. In such regions the lithosphere is thicker under the continent and thinner under the ocean. Last, lithospheric deformation in subduction zones occurs mainly under horizontal compressive stresses, as illustrated, in particular, by seismological evidence.

The next question that presents itself concerns the choice of rheologic models for the lithosphere and the underlying mantle. Numerous data, including evidences for plate rigidity and experimental deformation experiments [Goetze and Evans, 1979; Kirby, 1980, 1983] allow the assumption of an elastico-plastic lithosphere as a first approximation [Lobkovsky and Sorokhtin, 1976; Lobkovsky and Shemenda, 1981]. It is clear that the quantitative parameters of this lithosphere model, the yield limit \( \sigma_s \) and the modulus of elasticity \( E \), must be different from those derived from experiments on rock deformation that do not take...
into account scale and time factors. These parameters must have some effective values which can be found indirectly from various geologic and geophysical phenomena associated with lithospheric deformation (in particular, gravity anomalies).

The choice of a rheologic model for the asthenosphere is governed by the fact that its effective strength or viscosity is several orders less than that for the lithosphere. For this reason, one can use a fluid of very low (even zero) viscosity to model the asthenosphere (the part played by the asthenosphere is essentially to make Pascal's law hold beneath the lithosphere, i.e., to maintain hydrostatic equilibrium under the lithosphere).

The last question to be resolved is the driving force of subduction to be prescribed. The main sources of plate motion are considered to be slipping of the lithosphere from mantle highs in mid-oceanic ridges, action on the lithosphere of mantle (asthenosphere) currents which produce small tractions stresses at the lithosphere base acting along the currents, and gravitational sinking of the lithospheric slab into the mantle in the subduction zones.

The first two mechanisms can readily be replaced by local application of lateral compression to the slab using a rigid piston. This load should be applied far enough from the subduction zone (or its place of origin) for the boundary effects related to the loading method to be negligible.

The third driving mechanism of subduction (the slab being pulled into the mantle under its own weight) is modeled elsewhere [Shemenda, 1985, 1986].

Thus subduction is modeled in this work on a simple scheme involving an elastico-plastic model lithosphere overlaying a very low viscosity fluid, subjected to horizontal compression by means of a piston (Figure 1).

**SIMILARITY CRITERIA**

Modeling of subduction within the framework of the above scheme should satisfy the following similarity criteria [Shemenda, 1983]:

\[ \frac{s_0}{(pgH)} = \text{const}; \frac{E}{(pgH)} = \text{const}; \frac{p}{\rho_a} = \text{const}; \frac{Vt}{H} = \text{const} \quad (1) \]

where \( s_0 \), \( E \), \( H \), and \( \rho_a \) are the shear yield limit, the modulus of elasticity, thickness and density of the lithosphere, respectively; \( \rho_a \) is the asthenosphere density; \( V \) is the slab velocity; \( t \) is the time; and \( g \) is the acceleration due to gravity. Neglected here are Poisson's ratio in the lithosphere which is unimportant in our case and the pressure of ocean water on the plate surface. It should be noted that the last condition in (1) has a purely kinematic meaning and serves to convert model time into geologic time. The condition imposes no restriction on the process rate in the model.

**MODEL MATERIALS**

Substitution of the parameter values typical of the prototype (nature) into (1) leads to conclusion that the yield limit of the model lithosphere \( s_0 \) for the values of \( p \) and \( H \) that are realistic under laboratory conditions must be very small, of the order of 10 Pa. A material having this strength flows under its own weight and breaks when held in the hand. This is one of the main difficulties in the way of modeling large-scale tectonic processes; it does not permit one to make specimens of model lithosphere of required shapes and to perform mechanical operations with them before and after the experiments. The difficulty can be obviated by using special model materials. The lithosphere is modeled here by using a specially developed system of composite materials consisting of alloys of solid hydrocarbons (paraffins, 7.1%, and ceresins, 8.8%), mineral oils (61.1%), finely ground powders (23%), with a small addition of surface-active substances [Shemenda, 1981, 1984, 1989c]. In structure, this system is thixotropic dispersions of solid hydrocarbons and powders in oil. It possesses elastico-viscoplastic properties that are strongly dependent on the temperature; see Figure 2. This model material, as well as real rocks, exhibits different deformation properties in different ranges of temperature \( T \) and strain rate \( \dot{\varepsilon} \), from linear viscous to nonlinear viscous to plastic and brittle-dilatant properties. To model the lithosphere within the framework of the problem formulated above, these parameters were chosen in such a way that the yield limit \( s_0 \) satisfies similarity criteria (1) while being weakly dependent on \( \dot{\varepsilon} \) to model a plastic solid. The \( \tau(\varepsilon) \) curves in Figure 2d were derived in the range of \( \dot{\varepsilon} \) used in the model experiments. They demonstrate really a minor dependence of \( \tau_0 \) on \( \dot{\varepsilon} \) in the plastic flow regime. Also, \( \tau_0 \) almost does not depend on the confining pressure.

One more requirement which has been taken into account in fabricating the model materials was that they should be sufficiently hard and solid at room temperature, while the above properties should exist at a slightly increased working temperature of the model experiments. The plots in Figure 2d characterize the model material at the working temperature 39.5°C, where it has a very low yield limit. However, the material is a sufficiently strong solid at room temperature, so that one can perform the necessary mechanical operations on it without causing breakage.

The most convenient and suitable material for modeling the asthenosphere within the framework of the simple scheme described above proves to be pure water. Its density is \( \rho_w^0 = 10^3 \) kg/m³. The model lithosphere has the same density \( \rho_l^0 \) (it depends on the concentration of powder).

![Fig. 1. Scheme of the model experiments and the experimental installation: 1, bath; 2, "lithosphere"; 3, electric heaters; 4, piston; 5, "asthenosphere".](image)
EXPERIMENTAL INSTALLATION AND MODELING TECHNIQUE

To perform an experiment, water imitating the asthenosphere is poured into bath 1 (Figures 1 and 3) of dimensions 40 x 8 x 20 cm. A model lithosphere plate is placed above. The plate specimen is in direct contact with the bath walls, but special measures are taken to reduce the friction between the plate and the bath walls to nearly zero. Plane electric heaters 3 are used to make the entire system thermostatic at the working temperature 380-420°C (39.5°C for most experiments described below). The process lasts about 2 hours and is monitored using thermocouples which can be inserted into the plate itself. Piston 4 is then used to produce compression of the slab. The experimental installation ensures a constant level of the "free" mantle (water) top during the experiment. Plate deformation is studied through the transparent bath walls and by using a special technique for visualizing the deformation. The technique is as follows. The specimen is frozen before the experiment and then cut longitudinally at the middle into two parts. The section of a part is then treated with a special stamp to impress a grid matrix with lines 0.1 mm wide at intervals of 2 mm which is filled up with a paint suspension afterward. The two halves are then re-attached together. The resulting specimen is placed into the installation and deformed after heating. The experiment (compression) is stopped at the required stage. The model is then allowed to cool. The solidified specimen is removed from the installation and cut again along the previous section. The grid is then measured to derive the finite deformation and to identify faults. Circles 3-5 mm in diameter are also stamped into the slab along with the grid. Their deformation can readily be used to derive the directions of the principal strain and stress axes.

The normal horizontal nonhydrostatic (deviatoric) stresses $\sigma_h$ acting in the "lithosphere" at various stages of the experiment were also determined. These stresses cannot, unfortunately, be measured directly, because they are very small. For this reason the following procedure was used: a series of experiments was conducted under the same conditions. Prior to the experiments,
the model lithosphere was thinned at a certain location from below (a cut whose width was of the order of the specimen thickness \( H \) was made in the slab base all the way across it perpendicular to the direction of loading). The thinning was made ever larger from experiment to experiment, until failure occurred at the weakened location. Knowing the greatest thickness \( h \) of the thinned location where failure occurs, allowed the determination of \( \sigma_0 \): 
\[
\sigma_0 = \sigma_i \times h/H,
\]
where \( \sigma_0 \) is the yield limit at normal load (in that case a plane deformation takes place, and the relation \( \sigma_0 = 2\sigma_i \) is valid [Kachanov, 1969]).

\[ X_s \varnothing = 1.3 \times 10^5 \] m; \( p_m = 10^3 \) kg/m\(^3\). Conversion of the parameters into limit at normal load (in that case a plane deformation takes place, and the relation \( \sigma_0 = 2\sigma_i \) is valid [Kachanov, 1969]).

Deforation of the overriding (island arc) plate is accompanied by its frontal parts subsiding into the "trench" and by the formation of an archlike rise somewhat to the left, which can be identified as the frontal arc.

**Experiment 3.** Horizontal compression is applied to a lithosphere model of the transition zone between ocean and continent (Figure 6). The model consists of two parts of different thicknesses. The thicker part models the continental lithosphere; the thinner one models the oceanic plate. At first, uniform compression of the thin lithosphere results in the development of surface reverse faults as in experiment 1 (Figure 4). The "oceanic" lithosphere then loses stability and buckles (Figure 6b). A rise appears at the zone of thickness contrast (i.e. at the "continental margin"). Farther oceanward (to the left) a sagging occurs (Figures 6b and 6c). Continued compression increases the amplitude of buckling and reduces the associated wavelength. All deformation then concentrates in the sagging. Here, as a result of violent plastic compression, the specimen thickens (a "bead" forms) with accompanying thrusts formation and thickening of material in the top of the slab (Figure 6e-6f). This produces a fractured ridge overlying a more massive "lithospheric" root (Figures 6f and 7b). Localization of the deformation gradually creates nearly orthogonal shear (thrust) zones which cross the slab from top to bottom (Figures 6f and 7b). Further compression produces complete failure of the "lithosphere" along one of these zones and initiates subduction (Figures 5d, 6e, 6g, and 7c).
Fig. 4. Result of experiment 1. Horizontal compression of the oceanic lithosphere model. (a)-(f) Stages of the model process (for explanations see main text).

Fig. 5. Result of experiment 2. Compression of a plate containing a thinned zone.
Localization of deformation in the "oceanic" plate depends on the parameters of the "lithosphere", particularly, on the yield limit $\tau_s$ of the plate and its brittleness (i.e., on the magnitude of plastic deformation $\Delta e$ prior to failure; see Figure 2a). For instance, increased $\Delta e$ leads to stronger plastic crushing of the specimen in the zone of localized deformation and hence to a broader ridge and "lithospheric" root. Complete failure then occurs after great deformations. Roughly the same consequences follow from decreasing $\tau_s$. Also, along with decreased $\tau_s$, one notes a smaller bending amplitude $w$ of the oceanic slab (i.e., deformation is localized with failure occurring at smaller $w$). On the other hand, increasing the yield limit modifies the failure mechanism. The localization of deformation again occurs in the sagging but concentrates at the edges rather than at the center. Shear (reverse faulting) zones form there, preceding complete failure of the plate (Figure 7d). A subduction zone is then initiated at one of these. Similar results follow from decreasing the thickness of the "oceanic lithosphere" when $\tau_s$ is held constant.

Intermediate cases are also possible in which the bead (localization of deformation) and shear (thrust) faults form at the center and on the flanks of the sagging areas simultaneously at an early stage. Later, one of the mechanisms begins to dominate.

It is interesting to note that the pattern of model deformation obtained in experiment 3 (Figure 6) practically persists, if the "continental" and "oceanic" parts of the specimen are separated with a vertical cut beforehand. The only difference is that there is some vertical offset between the blocks along the cut in the latter case, owing to the fact that a thick slab under compression experiences a smaller uplift than a thin one. If, on the other hand, an inclined cut separates the slabs, subduction can occur along that cut. The experiments have shown that there is a critical value for the angle of inclination close to 90° which if exceeded, subduction is initiated at the cut.

The above procedure for determining horizontal stresses in the model lithosphere shows that experiment 3 (Figure 6) involves horizontal compressive deviatoric stresses $\sigma_h$ which are close to the yield limit $\sigma_s$ at the stage when a buckling instability arises in the "oceanic" plate. With further deformation, localization accompanies gradual failure of the material, and $\sigma_h$ gradually decreases, approaching a value of about 0.3$\sigma_s$ which is necessary to maintain the subduction.

Experiment 4. The "oceanic" plate is made longer and the "continental" one shorter in this experiment (Figure 8). Also, they are separated with a vertical cut. The result is similar to that in experiment 3. Experiment 4 clearly indicates a flexural
Fig. 7. Scheme of development of the plate localization deformation zone (from experimental evidence, experiments 3, 4, and 5): (a)-(c) stages of bead ("lithosphere" thickening) development; (d) another possible variant of the localization deformation and failure in the lithosphere (it is realized for diminished thickness $H$ and increased $r$); the zone of violent plastic compression (bearing strain) is shown in Figure 7a by the stippled area.

buckling of the "oceanic" plate. The localization of deformation develops as in experiment 3 in the sagging region closest to the continental margin.

Experiment 5. This experiment takes into account the weight of sedimentary material which may accumulate at continental margins. Such material exerts considerable pressure on the underlying transitional and oceanic lithosphere. Model "sedimentary" wedge (dotted in Figure 9a) of a special low-strength material having the following parameters when converted to nature: density $\rho^*_p = 2.5 \times 10^3$ kg/m$^3$; height of the base $h^*_2 = 6$ km; length $l^*_o = 150$ km (Figure 9) was placed on the oceanic slab. This slab separated from the "continental lithosphere" by the vertical cut, bends under the wedge weight. Similar to experiment 4, compression results in the "oceanic" slab loosing its buckling stability. However, in contrast to the preceding experiments in which a rise forms directly at the transition zone from "oceanic" to "continental" lithosphere, a depression developed at this location in experiment 5, initially enforced by the sediment load. In other words, the phase of slab flexure has been displaced by $\pi$. Accordingly, the depression in which deformation is localized and subsequent failure of the slab occurs, has shifted toward the "ocean" (Figures 9d and 9e). The same pattern persists in principle if there is no vertical cut at the "continental margin". The only difference is that in this case both "oceanic" and "continental" plates are involved in downwarping.

As a result the amplitude of the bending is reduced.

There is a critical value for the thickness of the sedimentary wedge $h^*_8$ below which the phase of slab flexure is not displaced. When $h^*_8$ has been diminished by a factor of 1.5-2 compared with experiment 5, the displacement no longer occurs. There is also some amount of "sedimentary" material at the continental margin in experiment 4 (Figure 8) but it does not exert an appreciable effect on slab deformation under compression.

It should be noted that the stress of the entire "oceanic" slab in experiments 3-5 is close to critical, that is, it is close to failure along its entire length. In some experiments with the conditions mentioned above, the "oceanic" plate experienced failure in different locations, including near the bath wall and directly in the transition zone under the "sedimentary" wedge (a fault formed there that was dipping under the "continent"). However, the most typical results were as described above (experiments 3-5).

Fig. 8. Result of experiment 4; this experiment differs from the previous by the longer length of the "oceanic" plate: in addition, the two plates are separated by a vertical cut.
The next three experiments (experiments 6-8) modeled subduction with an assigned inclined crosscut (fault) in the slab. The cut was made at a lower angle than the faults that formed in the preceding experiments. This was done to imitate later (well-developed) subduction stages for which the contact zone between the slabs is typically steeply dipping. Experiments 6, 7, and 8 differ by the values of the model lithosphere yield limit.

**Experiment 6.** In this experiment the slab has a relatively high $\tau_s^m$ value of 40 Pa. This leads to large vertical slab displacements in the subduction zone (Figure 10), producing an unrealistic trench depth which amounts to several tens of kilometers when converted to nature. The width of the trench is also too large.

**Experiment 7.** The model lithosphere has a low $\tau_s^m$ value of 1.2 Pa. In this experiment the range of vertical slab displacement proved to be too small (Figure 11). Also, the resulting trench is shallow, being the equivalent of only a few kilometers deep. Although the depth value is realistic, the trench is too narrow in this experiment. Its width is about 60 km when the scale factor has been taken into consideration, about half the width of real trenches. In addition, the distance from the "trench" to the "frontal arc" and the width of the outer rise proved too small.

Increasing $\tau_s^m$ compared to its value in experiment 7 improves the parameters characterizing the overall topography of the subduction zone, making them closer to the actual situation. For ex-
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Experiment 9 demonstrates the effect of the thickness of the subducting plate on the dimensions of the trench and the outer rise. The thickness of the "oceanic" plate is diminished in this experiment: \( H_m = 1.5 \times 10^{-2} \) m. This renders the amplitude of the outer rise greater, such that its elevation could be seen more distinctly (Figure 13). At the same time, outer rise became narrower as did the trench.

Last, we note the influence of the inclination of the plates contact surface on the topography of the subduction zone. Experiments have shown that a lower angle makes the trench shallower and the height of the outer rise smaller (Figure 14). The rise also grows in width. In addition, the frontal nonvolcanic arc becomes broader and lower.

**ANALYSIS OF EXPERIMENTAL RESULTS**

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Formation of a Subduction Zone in a Homogeneous Oceanic Lithosphere in the Transition Zone From Continent to Ocean

This scenario is the least trivial and perhaps the most general type of event. At first, increasing the horizontal compression of the lithosphere produces reverse faulting in its upper layer transverse to the direction of compression. Remembering that the upper layer in the actual lithosphere consists of brittle material of low strength, reverse faulting must be much more intensive in nature. Further, the slab loses buckling stability and assumes a wavelike shape with a wavelength $\lambda$ that depends on the plate parameters $H$, $\sigma_0$, and $E$. As the compression increases, $\lambda$ decreases. Prior to failure (localization of deformation) in the plate, $\lambda$ reaches average values of around $3H$ (i.e., about 200 km when converted to nature) and is weakly dependent on the other parameters of the "lithosphere". Afterward, the deformation is localized in one of the depressions (it may be initiated in several depressions, but continues to develop in one only). The lithosphere there thickens with the accompanying formation of a strongly dislocated ridge (Figures 6f and 7b) and two intersecting shear (in cross section) zones (Figure 7b). Failure along one of these results in the initiation of subduction. This mode of initiation of a subduction zone causes a well defined heterogeneous step like structure to develop on the inner trench slope forming after failure in the lithosphere (Figure 7c). The lower, steeper part of the slope comprises a strongly dislocated block of upper lithospheric rocks including oceanic sediments, splinters of crust, and upper mantle. The block forms the slope break dividing the inner slope into two different parts. In real environments a sedimentary terrace must form in front of the slope break which exerts a damming effect for the sediments coming from the island arc as shown in Figure 7c. Such features of structure do can be found along the inner slopes of some actual deep-sea trenches (see Figure 15).

The sequence of vertical seafloor movements in the frontal areas of the inner trench slope inferred from the experiments should also be noted, in particular, at the area of slope break. At first, when the slab loses buckling stability, this area subsides (Figure 7a). This is followed by uplift during the period of deformation localization (Figure 7b), and then by subsidence again when the trench forms and failure occurs in the slab (Figure 7c). The amplitude of these oscillations in nature must amount to a few kilometers.

It follows from the experiments that a localization of deformation must have been preceded by a buckling instability in the lithosphere. Such a phenomenon perhaps can be identified in the northeastern Indian Ocean. It deserves a more careful examination.

Intraplate deformation in the northeastern Indian Ocean. A seismic zone of intraplate deformation exists in this area within the Indo-Australian plate (see Figure 16). The deformation and seismicity are associated with extensive young NE thrusting (Figure 16) with which the observed seismicity is associated [Levchenko, 1986]. Fault density varies over the seafloor, but is particularly great in the northern Central Basin where reverse faults occur every 5-20 km [Levchenko, 1986]. Faulting is seen there superimposed upon the background of large-scale basement undulations (folds) which have a wavelength $\lambda$ of 100-300 km and amplitude 1-1.5 km [Weissel et al., 1980]; see Figure 17.

Fig. 13. Result of experiment 9 (the subducting plate thickness $H$ is as 1.5 times smaller than in the previous experiments).

Fig. 14. Influence of the dip angle of the interplate contact zone on the non-isostatic relief of a subduction zone. The thin horizontal line indicates the level of the isostatic equilibrium of the lithosphere; solid and dashed lines correspond to steeper and lower dip angles, respectively.

to the localization of plastic deformation in some part of the lithosphere. That process takes place when the compressive stress in the plate reaches the critical value, i.e., the yield limit for the lithosphere $\sigma$. Conversely, if there are weak zones in the plate or inclined faults all the way across the lithosphere (dipping at angles less than some critical value), then the subduction will initiate at the site of these features, because in that case initiation of subduction requires smaller compressive stresses. Below we consider variants.
Fig. 16. Indo-Australian plate: 1, subduction zone; 2, collision zones; 3 and 4, plate boundaries: 3, transform, 4, divergent [Weissel et al., 1980]; 5, epicenters of earthquakes with strike-slip focal mechanism; 6, thrust fault mechanism; 7, directions of the maximum compressional axis (from in situ measuremens) [Sykes, 1970; Stein and Okal, 1978]; 8, young (or rejuvenated) faults; 9, old transform faults [Levchenko, 1986]; 10, basement highs (pluses) and lows (minuses) [Weissel et al., 1980] (see inset); 11, location of the profile shown in Figure 17; 12, position of the hypothetical future subduction zone; 13, Komorine and 85 E ridges [Levchenko, 1986].

Fig. 17. Seismic and free-air gravity anomaly profiles across the intraplate deformation zone in the Central Indian Basin (the profile location is shown in the inset of Figure 16) [Weissel et al., 1980].
The flexure in the surface of the basement and the faults which cut it are largely flattened by the sedimentary cover. At the same time, the long-wavelength basement topography is clearly reflected in differential free-air gravity anomalies which reach Δg = 80 mGal there (Figure 17) [Weissel et al., 1980]. Such high values of Δg are difficult to explain by invoking isostatic compensation due, for instance, to crustal thickness variation. The basement topography is isostatically uncompensated, which is likely due to flexure of the whole thickness of the lithospheric plate [Weissel et al., 1980; Shemenda, 1989b].

The cause of the intraplate deformation seems to be subhorizontal northwest-southeast trending compression of the lithosphere, judging by ample seismological [Sykes, 1970; Stein and Okal, 1978; Wiens et al., 1986] and geomorphic [Weissel et al., 1980; Levchenko, 1986] evidences (see Figure 16). This compression of the Indo-Australian plate seems to have originated as a result of the collision between Hindustan and Eurasia [Weissel et al., 1980]. The deformation itself has been considered as a result of flexural buckling in the Indo-Australian plate due to the compression [Weissel et al., 1980; Cloetingh and Wortel, 1985]. The conditions for such buckling to occur have been studied by mathematical modeling [McAdoo and Sandwell, 1985; Zuber, 1987].

The overall pattern of deformation in the northeastern Indian Ocean is also fairly consistent with the buckling instability stage in the model lithosphere in experiments 4 and 5, Figures 8 and 9. The flexure wavelength λ in the model (with the scale factor taken into account) and in nature are consistent as well, being of the order of a few hundred kilometers (200 km on average). Also, λ is not constant along the oceanic plate in either nature (Figure 17) or the model (Figures 8 and 9) contrary to what one would expect based on an elastic model of the lithosphere. This is a result of the nonlinear plastic properties of the plate.

The model results are thus in essential agreement with actual situation. However, a more careful analysis of intraplate deformation and thrusting in the Indian Ocean raises a number of questions. In particular, the zone of flexure deformation is mainly bounded by the region adjoining the southern edge of Hindustan and Sri Lanka (Figure 16). This is not predicted by the model. In addition, recent studies [Levchenko et al., 1990] have established that the intraplate deformation zone is far from having a simple structure of linear depressions and rises striking northeast as previously supposed, mostly on the basis of gravity data [Weissel et al., 1980; Haxby, 1987], but rather is spotty: These questions may in part be related to the fact that our simple "two-dimensional" modeling has not adequately represented the "three-dimensional" nature of the situation in the zone of intraplate deformation, due to the complex geometry of the continental
below we present the result of an experiment which takes into account to a first approximation the configuration of the continental margin (i.e., the boundary between the thick and thin slabs) in the northern Indian Ocean and adjacent boundaries of the Indo-Australian plate.

Experiment 10. The northwestern Indo-Australian plate under study is a continuous layer in the model (Figure 18). The plate (layer) abuts a rigid stop (both wall) in the northwest, while the eastern and western boundaries of the plate, corresponding to the Sunda trench and the Carlsberg and Central Indian Ridges are free (the layer thickness along them is zero, the liquid "mantle" is at the surface). Compression is produced by a piston moving from the southeast at a constant rate (Figure 18a).

Similar to the previous experiments, the initial stage of deformation involves a plastic compression of the "oceanic lithosphere" which did not, however, occur uniformly over the entire area but with greater intensity in the region (approximately outlined by dots in Figure 18a) that abutted the Indian block on the south. Buckling stability in this region is lost on further compression. Two frontal folds which have a slightly curved shape around the "Hindustan" and decay toward the southwest, northwest and south, are most prominent. Instability develops approximately in the same way as in experiments 3, 4, and 5 described above (Figures 6, 8, and 9). Small reverse faults form on the plate surface, mostly in depressions (they cannot be seen in the photographs). The deformation then are localized in some places in the depressions, causing the formation of secondary strongly dislocated "localized" rises (ridges) which complicate the pattern of initial "flexural" folding. Later, inclined faults form traversing the entire lithosphere thickness. In contrast to the preceding experiments, these faults occur at different places and can be inclined in opposite directions. Figures 18a and 18b shows two such faults which have formed on the opposite sides of the "flexural" rise and dip in opposite directions (see Figure 18 caption). This "irregular" location of zones of deformation localization and faults is typical of three-dimensional models. Further compression produces thrust (underthrust) displacements along the new faults, with the faults themselves getting longer (Figure 18b). As they grow, slab instability (flexure) occurs on both sides of the fault zone (between it and the lateral plate boundaries). The folds which arise there are perpendicular to the compression axis and are not conformable with the trend of "older" folds in the central zone. Upon the background of fault growth and fold development, thrust-strike slip faults arise on the eastern side of the "Hindustan" at some instant of time (Figure 18c). Faults also then occur on the sides of the central zone in conformance with the trend of folds developing there. These faults interact with those propagating out of the central zone and produce the ultimate shape of a new convergent plate boundary the configuration of which is outlined approximately by a dashed line in Figure 18d. This boundary did not form in one event as a single zone of one-directional subduction but resulted from a sequence of restucturings in local subduction zones that formed at individual faults and had different polarities.

This three-dimensional modeling further confirms the existence of flexural buckling in the lithosphere of the Indian Ocean and elucidates some of the finer details in the structure of the zone of intraplate deformation. In particular, the experiments corroborate
the intuitively understandable cause of deformation concentrating in the northern Central Basin as being due to the continental Hindustan block jutting out into the oceanic lithosphere. Also, the model provides insight into the possible causes of the complex irregular (spotty) structure of the deformation zone. It may be related to rises of two types, those due to lithospheric flexure and due to the localization of deformation. Certain strongly dislocated basement highs that can be detected in seismic profiles traversing the onset of deformation can be interpreted as "localization" ridges.

**Polarity of subduction zones: trapped back arc basins.** Experiments 3-5 show (Figures 6, 8, and 9) that failure in the lithosphere due to horizontal compression can occur not directly in the continental margin, even though there are subvertical faults through the entire lithosphere, but at some distance toward the ocean. This rather surprising result is due to buckling in the oceanic plate prior to failure. The wavelength of the buckling controls the distance from the continental margin to the location of failure in the lithosphere.

Experiments have shown that complete failure of the "oceanic" plate in zone of deformation localization can occur along two intersecting planes dipping in opposite directions (Figures 6c and 7b). For example, the fault plane in experiment 3 (Figure 6) dips under the "continent", while in experiments 4 and 5 (Figures 8 and 9) it dips seaward. It has not been reliably established which direction is the preferred one. A large series of experiments would be necessary to come to a conclusion about this. Both possibilities seem equally likely as demonstrated by the results of the following experiment.

**Experiment 11.** Compression of a model lithosphere consisting of two parts having different thicknesses. The conditions of this experiment are the same as those of experiment 3 (Figure 6), except that the specimen of model lithosphere has first been separated into two halves by a longitudinal vertical cut. The initial stages of experiment 11 are similar to the experiments described above: the "oceanic" lithosphere bends as a whole and thrust faulting develops on the surface. After that a zone of deformation localization common to both halves starts to develop where, as usual, conjugate inclined thrust zones (faults) form. However, complete failure of the "lithosphere" in the two halves occurred along different shear (thrust) zones, i.e., along planes dipping in opposite directions. As a result, subduction began to develop in opposite directions too, which gradually displaced subduction zones on both sides of vertical fault (cut); see Figure 19.

Failure of the lithosphere along one or the other directions has important geotectonic consequences. For example, if the surface of the plate fracture dips under the ocean, as was the case in experiments 4 and 5 (Figures 8 and 9), then consumption in the resulting subduction zone of the part of oceanic lithosphere between the subduction zone and continental margin inevitably leads to their collision and the result is that the subduction zone is claimed. Consequently, a new failure of oceanic plate occurs directly at the continental margin, and the subduction settles down to proceed toward the continent. In this process, structures of the attached subduction zone happen to belong to forearc areas of the new overriding plate. A similar episode may have occurred during the evolution of the Andean margin of South America, this being the only convergent plate boundary where subduction goes on directly beneath the continent.

If, on the other hand, plate failure in the zone of deformation localization occurs along the fault dipping under the continent (as in experiment 3, Figure 6), then a trapped marginal basin separated from the nascent island arc and the continent by the arising subduction zone can exist for a long time. The dimensions of the basin are controlled by several factors. When the simplest "two-dimensional" case is considered in which sediment load at the continental margin is absent (experiments 3 and 4, Figures 6 and 8), the basin width d is half the wavelength of lithosphere flexure at the stage of formation of the buckling instability, i.e., is of the order of 100 km. The presence of a sedimentary wedge in the continental margin may significantly increase d (experiment 5, Figure 9). Also, the configuration of the continental margin may play an important part in determining the dimensions of the trapped back arc basins. This was graphically illustrated in experiment 10 (Figure 18). Local subduction zones originate first near the protrusions of the continental lithosphere into the ocean, and then propagate to meet each other along the shortest routes. Resulting subduction zone do not mimic the configuration of the continental margin.

Natural analogies of trapped back arc basin are the Aleutian Basin in the Bering Sea, the Sea of Okhotsk, excluding the Kurile trough and some others.

**Application of the model to oceanic transform faults.** A small compressive or tensile component is always present in oceanic transform faults along with the main strike slip one. The sign and magnitude of that component largely determine the relief and deep structure of the transform fault [Ushakov et al., 1979; Dubinin, 1987]. The model in Figures 7a and 7b can formally be applied to faults involving transverse compression. Indeed, plates of different ages and hence different thicknesses are in contact across a transform fault. If the transform fault is regarded as a two-dimensional feature, then strike-slip displacement on it can not produce lithospheric deformation. Deformation may be due to the normal displacement component, in particular, compression. The problem of lithospheric deformation in a transform fault due to such a compression is similar to that considered above (Figures 6, 8, and 9), if the fault itself is subvertical. According to experimental results a ridge should form by the localization of lithospheric deformation (including crustal deformation) at a distance of about 100 km from the fault. Verification of this hypothesis requires a detailed analysis of the relevant transform faults. This is not, however, a subject of this work. We merely mention that the "region of influence" of some transform faults
characterized by transverse compression does involve features (ridges) that can be regarded as being due to localization of a lithospheric deformation. The features in question are the Gorindge Ridge in the eastern Azores-Gibraltar fault [Verzhbitsky et al., 1989] and the seamounts of the Hossu system some 100 km southward of it. Another example is the northeastern part of the Owen transform fault in the Indian Ocean. There is also a rise associated with crustal thickening on the northwest of it at a distance of about 100 km [Whitmarsh, 1979].

It is supposed that northward subduction is already occurring under the Gorindge Ridge [Le Pichon et al., 1970]. This supposition is also consistent with the model. Subduction can indeed start around a transform fault, providing the amplitude of the transverse compression is large enough.

To summarize, analogies for all stages of failure occurring in the oceanic lithosphere under horizontal compression can be found in nature. The initial stage of the process, buckling instability developing in the lithosphere (Figure 7a), seems to take place in the northeastern Indian Ocean, the intermediate stage, localization of deformation with crustal thickening (Figure 7b), applies to the Gorindge Ridge, with the final stage illustrated by numerous subduction zones.

**Initiation of a Subduction Zone on an Old Dipping Fault**

There is another, simpler mode of formation of a subduction zone which can occur if an old dipping fault or weakened zone striking across the horizontal compression is present. In such situations, subduction on the dipping fault will be initiated, before the compressive stresses that are sufficient for buckling instability to develop in the lithosphere have been reached. The inner trench slope produced therefore will have a structure that is much different from that in the preceding model. It is largely controlled by a deformation at the frontal edge of the overriding wedge which forms when the edge caves down on the surface of the subducting lithosphere during initial stages of subduction. This effect can be seen in experiments 6-8 (Figures 10-12) where the originally horizontal overriding slab subsided above a fixed dipping fault at the beginning of the subduction process. The subsidence took place when the unsupported portion of the overriding wedge was heavy enough to overcome its yield limit \( \tau_s \) (the \( \tau_s \) of the wedge has not been exceeded in Figure 10 but has been in Figures 11 and 12), i.e., when conditions became available for plastic deformation in the wedge. The deformation in question was a superposition of flexure producing extension in the upper layers of the wedge and compression in the lower layers, combined with shearing (normal faulting) (Figure 20). Greater detail can be discerned in experiment 12 (Figure 21).

**Experiment 12.** A horizontal layer that possesses dilatant-plastic properties is placed on a rigid base consisting of two parts that could be displaced relative to one another in a subvertical direction (Figure 21). Such a displacement makes one area of the model lithosphere subside relative to the other. Although the displacement in the layer is subvertical, faults striking across that direction form at the initial stage of the subsidence (Figure 21a). Further development is dominated by normal faulting, with normal faults and grabens forming on the surface. At first sight, a puzzling situation emerges: in spite of the fact that the lithosphere as a whole is under considerable compression in the subduction zone (including the trench), both trench slopes are tensional features (Figure 20) (extension of the outer trench slope arises due to bending of the oceanic lithosphere, see below). This structure does occur in some regions, for instance, in the Central American (see Figure 22), Tonga-Kermadec trenches [Gnibidenko et al., 1985] and some others.

Returning to experiments 6-8 (Figures 10-12), one notes that it is only the most frontal parts of the overriding wedge descending below the initial horizontal level during subduction which experience subsidence. Farther up the slope, the wedge is on the contrary uplifted and forms an isostatically uncompensated frontal arc. The uplift is also accompanied by plastic deformation that produces reverse faults as shown in Figure 20.

The blocky structure of the overriding wedge which forms during the initial subduction stages continues to develop during subduction due to various factors. For example, reverse faulting can occur due to friction with the subducting slab on the originally normal faults that have formed in the forearc portions of the wedge [Lobkovsky et al., 1980]. Still greater changes can be associated with modified angle of dip for the contact zone between the slabs. The change in dip may arise due to a nonuniform rate of working out (tectonic erosion [von Huene and Lallemand, 1990]) of material at the base of the overriding wedge. If the dip becomes smaller (and, accordingly, the forearc region subsides, while the most frontal portion of the overriding wedge is uplifted, Figure 14), then the direction of movement...
along the faults in the overriding slab shown in Figure 20 will be reversed. Such effects can also occur when the pressure of the subducting slab on the overriding one is changed owing to changes in subduction from the Chilean to the Mariana types and vice versa [Shemenda, 1985].

Stresses and strain in the subducting plate. While the bulk of deformation and hence formation of the structure of the overriding wedge takes place when the subduction zone is initiated, the subducting plate experiences large quasi-stationary deformations at all time. These were studied by the technique for visualizing deformation described earlier based on curvature of the originally rectangular grid and of circles drawn on the model lithosphere. The technique incorporates the concepts and familiar solutions of plasticity theory. The subducting plate has been shown to experience flexure in front of the subduction zone, leading to compression of the material at the base of the plate and extension at the top (Figure 20). Normal horizontal stresses in the top and bottom of the slab across the trench have exceed the yield limit in the area involving the outer (oceanic) trench slope and part of the outer rise leading to plastic deformation there. The subducting slab has increasing curvature from the outer rise towards the trench during later stages of subduction. This leads to increasing plastic deformation and development of plastic zones inside the slab (dashed in Figure 20). The growth of the deformation is accompanied by small linear normal faulting on the plate surface in the outer trench slope parallel to the trench. Because the average stress in the plate during subduction is compressive in the cross section (about 0.3σs), the picture is asymmetrical in that the lower plastic zone is thicker than the upper one (Figure 20).

The flexure of the subducting slab becomes more complex closer to the trench where the slab curvature is at a maximum. In the zone of maximum bending marked as a stippled area in Figure 20, the thickening upper and lower plastic zones are merged and connected by sliding lines which accommodate shearing (reverse) faulting movements. When the region of maximum bending has been passed, the thickness of the plastic zones and the deformation in these decrease sharply, the slab then sinking into the "mantle" without further deformation. It should be emphasized that this situation is obtained because the subducted portion of the slab is in a hydrostatic equilibrium with the surrounding "mantle" and does not experience any dynamic influence due to it. If there is no equilibrium, the deformation pattern would apparently change.

To summarize, the comparatively sharp change in the direction of movement of the subducting plate in the subduction zone occurs, not only owing to extension and contraction of plate material due to flexure but also due to shearing (thrusting) motion. The significance of the shear in terms of total slab deformation decreases with decreasing dip angle, ψ, of the contact surface between slabs. The zone of shear (thrust) deformation becomes extended and vanishes altogether in the limit. In contrast, increasing ψ as well as decreasing τs makes shear deformation more intensive such that shear zone becomes narrower and better defined in outline as is seen in the next experiment.

Experiment 13. Subduction occurs under a stiff stop dipping at an angle of 45° which occupies the place of the overriding slab. The subducting plate experiences a sharp flexure in front of the stop. Its internal deformation can be inferred from Figure 23. The character of curvature in the transverse lines shows that shear (reverse) faulting is dominant, developing around the line AB in Figure 23. To see this, one should compare the experimental pattern (Figure 23) to the ideal scheme (Figure 24) of pure shear derived by simple geometrical constructions. Although the patterns are roughly similar, the experimental result is more complicated. The causes of this are, first, that the shear zone is not straight and, second, that practically the entire flexure zone involves comparatively small deformation. Two other zones can be identified upon the background of this deformation: one lies in the lower part of the inflexion region and the other (ACD, Figure 23) under the stop (appropriate sliding lines are shown in Figure 26). The lower zone dominating the slab base is associated with flexure and compression while the upper zone results from the fact that it is the upper layers of the slab which are the first to experience the pressure exerted by the stop. They are, as it were, pressed into the body of the rest of the specimen. In addition, friction exerted by the stop acts as a brake on the top of the slab (the value of friction is related to angle γ).

Fig. 22. Schematic section through the Guatemala trench [after Aubouin et al., 1982].

Fig. 23. Result of experiment 13. Subduction under a rigid stop dipping at an angle of 45° (the part of the deformed and frozen specimen in the subduction zone is shown in the photograph; the transverse lines in the plate had originally been vertical and perpendicular to the surface of the undeformed specimen; dashed lines show the main sliding lines (see the text for more explanations).
When subduction develops at a steeper angle, the shear zone does degenerate into a straight line in the experiments, as shown in Figure 24. This is easily seen in experiment 1 in Figure 4. The existence of such a line (zone) and the major role played by shear movements in the oceanic plate in the subduction zone have been hypothesized by Lobkovsky and Sorokhtin [1976].

Thus the width of the region of shear deformation in the subducting slab, its position relative to the trench, and the intensity of deformation in it depend on the specific subduction conditions. It can be expected that when the shear (thrust) deformation in the region of maximum flexure is large enough and concentrates along a comparatively narrow zone (as in Figure 23), it should be detectable in the distribution of earthquake hypocentres. It is possible that the hypocentral swarm observed in the distributions of earthquakes for some subduction zones which form the focal zone dipping oceanward (Figure 25), have just this origin.

Estimation of Departure From Isostasy in the Subduction Zone

Analysis of the experimental results (Figure 23) allowed the determination, at least in part, of the structure of the field of sliding lines in the region of a plastic deformation (Figure 26) and the conclusion that the plastic zone ACD abating the stop (overriding plate) is under uniform stress. This allows derivation of the contact nonhydrostatic pressure $\sigma_n$ which the subducting plate exerts on the overriding plate using plasticity theory. This pressure has proved to be $\sigma_n = -0.9\sigma_t$ [Shemenda, 1979, Lobkovsky and Shemenda, 1981]. The full pressure $P$ at each point of AC can be derived by adding the term $\rho g z$ to $\sigma_n$ ($z$ is the vertical coordinate of the point).

The mean pressure $\rho g z$ corresponds to the full hydrostatic equilibrium at depth $z$. The excess $\sigma_t$ disturbs the equilibrium, producing some nonisostatic uplift in the overlying layers (those above the plate contact). The friction between the plates, $\tau_n$, tends on the contrary to pull them downwards. The net vertical pressure $\sigma_v$ is: $\sigma_v = \sigma_n + \tau_n g \sin \phi$. The friction $\tau_n$ in experiment 13 can be estimated from the value of $\gamma$, the angle which characterizes the upper zone of uniform stress in Figure 23 as $\tau_n = \tau_0 \cos 2\gamma$ [Hill, 1950]. Assuming the pressure $\sigma_v$ to be fully spent on producing a piston uplift of the lithosphere areas lying directly above the line AC (Figures 23 and 26), one can determine the uplift amplitude $\Delta H$: $\Delta H = \frac{-\sigma_v \rho g}{\rho g} = \tau_0 (0.9 - \cos 2\gamma \sin \phi)/\rho g$. Substituting the values $\tau_v = 1.3 \times 10^9$ Pa, $\psi = 42^\circ$, $\gamma = 22^\circ$ (see Figure 23), we get $\Delta H = 1$ km. Thus, there must be a free-air gravity anomaly $\Delta g$ above the nonisostatically uplifted layer of thickness $\Delta H$. The anomaly can be estimated by using the formula for a plane parallel layer [Lobkovsky and Sorokhtin, 1976]: $\Delta g = 2\pi f_0 \Delta h = 140$ mGal, where $f = 6.67 \times 10^{-7}$ m/(kg x s$^3$) is the gravitational constant. Positive gravity anomalies of this order are typical of actual subduction zones. If the interplate friction $\tau_n$ in the subduction zone were absent ($\gamma = \pi/4$), then the gravity anomaly would equal $\Delta g = 470$ mGal.

Elastico-plastic Bending of the Lithosphere Plate and Estimation of Its Yield Limit and Modulus of Elasticity

The excess pressure between the plates leads on the one hand to a nonisostatic uplift of frontal parts of the overriding lithosphere and on the other to bending of the subducting plate and formation of a deep-sea trench. On account of the elastic properties of the lithosphere and the presence of an effectively liquid base beneath it, plate deformation due to bending is transmitted over a considerable distance seaward, creating an extended rise of small amplitude known as the outer rise. This subduction-related phenomenon is well known, such that there is little doubt that the trench and the outer rise are caused by quasi-elastic bending of the subducting plate. Gunn [1947] was the first to suggest a mathematical model for this bending. Numerous later studies were largely concerned with refinements of his model in order to
derive theoretical solutions that better corresponded to actual bathymetric profiles across various subduction zones. Such solutions were used to evaluate parameters of the lithosphere. This approach encounters several difficulties. They are related to an indeterminacy of the boundary conditions for the bending plate in the subduction zone and the nonstationarity nature of this bending at the earliest stages of subduction (only these stages can usually be modeled mathematically). The relief at these stages, as shown by experimental results, is considerably different from that at later subduction stages. As shown in Figure 10b, the maximum bending of the subducting plate at the beginning of subduction occurs in the upper part of the outer trench slope (see the arrow in Figures 10b and 10c), while practically no bending is observed in the lower part. An equilibrium (steady state) shape of the bending is established during subduction, with the part having maximum curvature being shifted trenchward (Figure 10e). The trench becomes shallower and narrower in the process.

Certain difficulties also arise when experimental modeling is used. For example, a comparison of experiments 6 through 8, Figures 10 to 12 yields a fairly reliable estimate of $\sigma_s$, as the plate shape in the subduction zone strongly depends on that parameter. The best fit is obtained in experiment 8 (Figure 12) for $E_0 = 2\sigma_s = 30$ Pa, as mentioned earlier. Adopting $H = 60$ km as lithosphere thickness, we use (1) to get a value of $\sigma_s = 2.6 \times 10^8$ Pa. Similar values have been derived by other workers from gravity data [Lobkovsky and Sorokhtin, 1976; Usakov, 1968].

The effective modulus of elasticity $E$ for the plate has also been varied. However, the changes in the deformation of the lithosphere were too small to be observed. Therefore numerical modeling of the initial stage of subduction has been used within the framework of formulation which is similar to the initial stages of experiments 6-8 (Figures 10-12) [Tishchenko, 1985]. The value of $\sigma_s$ was fixed to be that which followed from the experimental modeling. The results indeed show a weak dependence of the shape of the bending plate on $E_0$. Any of the values of $E$ within range $10^{10}$ to $10^{11}$ Pa are suitable. Outside of that range, the amplitude of the outer rise is either too small or too large. The value which was used in the experiments described here lies within the above range. The values of lithosphere parameters adopted at the start of this work $\sigma_s = 2.6 \times 10^8$ Pa and $E_0 = 1.7 \times 10^{10}$ Pa thus follow from modeling results. At the same time they are consistent with the results of other works.

It is interesting to note that the width of the outer rise is 300 to 400 km on average. The plate bending wavelength $\lambda$ is then 600 to 800 km. On the other hand, the buckling instability for the lithosphere in the zone of intraplate deformation in the Indian Ocean has a wavelength that is smaller by a factor of three or four. The difference seems to be due to a different horizontal compression of the lithosphere for both these cases. With small or no compression, $\lambda$ reaches values of about 1000 km, while when the compression is close to $\sigma_s$, the value of $\lambda$ diminishes to a few hundred kilometers. Experiments have been carried out in which the subduction zone was "jammed" at some stage of subduction (this was done by local cooling of "lithosphere" material in the contact zone between the plates). Since subduction became impossible, the motion of piston 4 (Figure 1) led to a greater compression of the model. This was accompanied by a considerable contraction in the length of the outer rise and a growth of its amplitude. At the same time the flexure (sagging) on the seaward side of the rise increased. Subsequent compression produced a localization of deformation there in the sagging (similar to in experiments 3-5, Figures 6, 8, and 9), the plate experienced failure then and a new subduction zone formed.

**CONCLUSIONS**

1. Compression of lithosphere in the area of a passive continental margin produces a buckling instability in the oceanic plate having wavelengths of a few hundred kilometers. Later, a localization of deformation occurs in a downwarp at some distance from the margin, with accompanying thickening of the lithosphere and formation of a linear ridge due to the thickening and thrusting. The plate then experiences failure along the dipping zone, and subduction starts. The inner trench slope which forms during failure of the lithosphere has a typical scarped structure. Its lower part contains a dislocated block of oceanic crust and sediments forming the slope break. In front of this there is a "reservoir" which is filled with arc-derived sediments to form a deep-sea terrace. Some actual trenches have a similar structure.

2. Different variants of the above failure mechanism for a continuous lithosphere are possible if it contains zones where the lithospheric thickness or strength is reduced.

3. Modeling of the situation existing in the zone of intraplate deformation in the northeastern Indian Ocean corroborates an earlier hypothesis of a buckling instability of the lithosphere developing in the area, and of the initiation of a new subduction zone.

4. If there is an old dipping fault striking across the compression of an oceanic lithosphere, it is on that fault that a subduction zone is initiated. The inner trench slope has a different structure and forms due to normal faulting which develops in the frontal part of the overriding plate. The formation of a subduction zone then requires less compression than in the preceding case by a factor of 2 or 3.

5. In contrast to the overriding plate where deformation develops mainly during the initiation stage of subduction, the subducting plate experiences continual large quasistationary deformations. These are controlled in front of the subduction zone by an elastico-plastic bending of the oceanic lithosphere that increases closer to the trench. The bending becomes so large directly in the subduction zone (under the overriding plate) that shear (thrust) faulting can develop (depending on specific conditions) along the zone dipping under the ocean and crossing the entire lithosphere.

An analysis of the stress-strain state in the model subducting plate yields the pressure exerted by the subducting lithosphere on the overriding plate and permits an estimation of the free-air positive gravity anomaly due to it. The anomaly amounts to a few hundred milliGals.

5. The best agreement between the generalized relief of the subduction zone in the model and nature is achieved when the following effective values are adopted for the real lithosphere: shear yield limit $\tau_s = 1.3 \times 10^8$ Pa; modulus of elasticity $E$ about a few times $10^{11}$ Pa, and thickness $H = 60$ km.

**NOTATION**

- $\sigma_s$: lithosphere yield limit under normal loading.
- $\tau_s$: lithosphere shear yield limit.
- $\sigma_h$: horizontal nonhydrostatic stress in the lithosphere.
- $E$: lithosphere modulus of elasticity.
H  

lithosphere thickness.

h  
thickness of thinned segment of the plate.

h₀  
thickness of the sedimentary wedge on the passive continental margin.

l  
length of the sedimentary wedge.

d  
width of the marginal basin.

Δhₙ  
nonisostatic vertical displacement of the surface of the overlying plate.

ρ₁  
lithosphere density.

ρ₀  
aethosphere density.

g  
acceleration due to gravity.

f  
gravitational constant.

τ  
shear stress.

e  
deforestation.

ε  
strain rate.

τₙ  
fraction between plates in the subduction zone.

σₐ  
nonhydrostatic pressure between plates in subduction zone.

σᵥ  
vertical pressure in the interplate surface.

T  
temperature.

λ  
dip angle of the interplate surface in the subduction zone.

γ  
angle characterizing plastic deformation in the subducting plate.

Δg  
free air gravity anomaly.

V  
plate velocity.

t  
time.

w  
deflection of the lithosphere surface from the isostatic equilibrium level.

α,β  
families of sliding lines.

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