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TECTONIC STRESS - MODELS AND MAGNITUDES

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Introduction

Understanding the state of stress in the Earth's lithosphere is one of the paramount problems in Earth tectonics. The stress state is linked to causes - loading and unloading, heating and cooling, plate motions and driving forces, to consequences - creep deformation and seismic failure, and to rheology - the depth over which stress can be supported and the time dependence of material properties. None of the classes of links has been characterized in sufficiently quantitative detail to define the stress tensor in the lithosphere without ambiguity and without a long inference chain involving poorly tested assumptions. This paper deals with one cause of stress in the lithosphere - the system of forces that maintain plate motions. Specifically addressed are ways by which models of tectonic stress in the plates can be used to constrain the magnitude of regional deviatoric stress in the Earth's lithosphere.

Global models of the intraplate deviatoric stress that arises from the driving and resistive forces controlling plate motion have been given by Solomon et al. (1975) and Richardson et al. (1976, 1979). A principal objective of those studies has been to find those sets of forces that best match the body of intraplate stress observations. The observations felt most reliable for such comparisons, however, are the indications of principal stress directions inferred from the mechanisms of midplate earthquakes, from in situ stress measurements, and from the strikes of stress-sensitive geological features. While a comparison of model predictions and observations on the basis of principal stress orientations is straightforward and serves as a useful test by which to reject possible force models, such an exercise does not directly address the absolute <u>magnitude</u> of intraplate deviatoric stresses, since all deviatoric stresses in a model can be multiplied by an arbitrary constant without changing the relative magnitudes or the orientations of the principal stresses. We show in this paper, however, that under certain conditions the body of data on intraplate stress orientations <u>does</u> constrain the magnitude of tectonic stresses.

It might be argued that stress magnitudes can in principle be measured by direct in situ techniques in sufficient locations to characterize the stress field for length scales comparable to plate dimensions, thus obviating the need to apply indirect arguments to constrain tectonic stress magnitudes. This eventuality is doubtful for the near term, because of the difficulty in interpreting near-surface measurements in terms of lithospheric characteristics, and because further advances in technology will be necessary to conduct routinely measurements of in situ stress over the large fraction of the Earth's surface covered by oceans. These difficulties are discussed at length in other papers of this volume.

Thus the question remains: given the large and growing body of data on the orientations of principal stresses within the plates, what information on the magnitudes of regional deviatoric stresses can be obtained from numerical models for tectonic plate stresses? We discuss in this paper two routes by which useful information on stress magnitudes can be derived: (1) For the driving force models that best fit the stress orientation data, if independent information on the magnitude of one or more of the forces in the system can be obtained, then the magnitudes of the total predicted stress field are constrained to comparable precision. The best fitting force models we have examined to date all involve a significant contribution from ridge forces, the pushing forces that arise because of the elevated topography of ridge axes with respect to abyssal sea floor. Since ridges exert forces equivalent to compressive plate stresses of 200-300 bars magnitude, this leads to the prediction that regional deviatoric stresses are of this magnitude. (2) If in the vicinity of a known local source of stress, the observations of stress orientations indicate comparable control by the local and regional stress field, then the magnitude of the regional field may be estimated. This line of argument holds special promise for oceanic intraplate regions where earthquakes have occurred in the vicinity of islands or large bathymetric features characterized by sufficiently good topographic and gravity data to model the associated local lithospheric stress.

It should be mentioned that when direct in situ measurements of stress magnitudes have high reliability, the magnitude data can be used alongside the stress orientation data as a more powerful set of constraints on both regional and local forces on the lithosphere.

Stress Magnitudes and Global Plate Models

The comparison of predicted and observed directions of principal intraplate stresses can be a sensitive test of possible sources of stress. As noted above, if such a comparison indicates a significant contribution from a source of stress of known or estimable magnitude, then a strong constraint on the general magnitude of deviatoric stress in the lithosphere on regional scales is obtained. In this section, we summarize our recent work (Richardson et al., 1979) on testing global models of intraplate stress predicted by plate tectonic driving forces against observed directions of principal stresses, with particular emphasis on possible inferences on the magnitude of deviatoric stresses.

Premises. That observations of principal stress directions in the plates can be used to constrain plate tectonic driving force models requires the adoption of three working premises: (i) that regionally consistent stress orientation fields exist for large fractions of the stable interiors of plates, (ii) that such stress fields are steady over time periods less than that ($\sim 10^6$ years) characterizing changes in plate motions; and (iii) that a recognizable portion of these stress fields are dominated by contributions from plate tectonic forces.

The first premise has substantial basis in fact for most of the plates (Sykes and Sbar, 1974; Sbar and Sykes, 1973; Richardson et al., 1979); see Figure 1. The second premise depends on the question of whether in plate interiors the deformation and stress arising from past plate boundary slip superpose to produce steady motion and stress, or whether individual stress 'waves' from large earthquakes are discernible (e.g., Anderson, 1975). This issue will be addressed by ultra-precise geodetic measurements of short-term plate motions soon to be made (Niell et al., 1979; Smith et al., 1979; Bender et al., 1979). The third premise will be the most difficult to establish with certainty, but is a reasonable working hpyothesis for regions well removed from such other notable sources of stress as recent tectonic or thermal activity, recent topographic loading or unloading, and pronounced structural heterogeneities.

Possible Driving Forces. We consider several simply parameterized driving and resistive forces as potential elements of a plate tectonic force model: plate boundary forces at ridges, trenches, transform faults and zones of continent-continent collision; and basal forces associated with viscous interaction between the lithosphere and the asthenosphere. While all of these forces contribute to lithospheric stress, it is important to recognize that potentially large stress contributions can also arise from lithospheric cooling (Turcotte and Oxburgh, 1973), latitudinal plate motion (Turcotte and Oxburgh, 1973), crustal thickness inhomogeneities (Artyushkov, 1973), lithospheric loading and unloading (Walcott, 1970; Watts and Cochran, 1974; Haxby and Turcotte, 1976) and ancient tectonic events (Swolfs et al., 1974; Tullis, 1977). Care must be excerised in the interpretation of stress observations in terms of plate forces to remove or to avoid where possible the effects of these additional sources of lithospheric stress.

The compressive stress exerted on oceanic plates by the elevation of mid-ocean ridges is the easiest to quantify among the set of possible driving and resistive forces, and is in the range 200-300 bars (Hales, 1969; Frank, 1972; McKenzie, 1972). At subduction zones, the negative buoyancy of subducted lithosphere is capable of exerting extensile stresses of several kilobars stress on the adjacent plates (McKenzie, 1969; Turcotte and Schubert, 1971), but the greater fraction of available pulling force is counterbalanced by forces resisting descent of the slab into the mantle (Smith and Toksöz, 1972; Forsyth and Uyeda, 1975; Richter, 1977). The net pull by slabs on the surface plates is uncertain but is considerably smaller than that due to available negative buoyancy. At zones of continent-continent collision, the net force on the adjacent plates may be resistive (net compression), because of the contribution from the excess topography of the mountain belt marking the collision zone; the contribution from topography involves shear stresses of 200-300 bars for the main boundary fault at the base of the Himalayas (Bird, 1978). The resistive force at transform faults is among the least certain of plate boundary forces (Brune et al., 1969; Brace and Byerlee, 1970), but is not likely to be a major contributor to the plate driving mechanism on the basis of the relatively small fraction of boundary taken up by transforms for most plates and the poor correlation of plate speeds with length of transform boundary (Forsyth and Uyeda, 1975; Aggarwal, 1978).

The viscous traction at the base of the plates is uncertain both in magnitude and in direction. The uncertainties are linked to questions of the radial scale for upper mantle convection, the planform for 'counterflow' to balance plate creation and destruction, and the existence of a smaller secondary scale of asthenospheric convection to transport heat (Richter and Parsons, 1975;

McKenzie and Weiss, 1975; Chase, 1979; Harper, 1978; Hager and O'Connell, 1979). Some simple forms for viscous drag are adopted as a basis for testing models, but the various potential complexities must be kept in mind.

Stress Models. A variety of driving force models incorporating different relative amounts of boundary forces and basal tractions as described above have been tested against the observations of intraplate stress orientations. The lithosphere is modeled as a thin, spherical, elastic shell, and stresses are calculated from the imposed forces using the finite-element analysis described by Richardson (1978). The results of many models are given in Richardson et al. (1979), and only a summary of the results pertinent to the question of stress magnitudes will be given here.

A summary of stress orientation data for intraplate regions is given in Figure 1. Most of the data come from the mechanisms of intraplate earthquakes; a lesser number of data are from in situ measurements (see Richardson et al., 1979, for the original sources of the data shown). An important question for such a data set often is which data to include and which to exclude as constraints on the tectonic stress field. While the answer to such a question is of necessity at least partly arbitrary, our approach has been to exclude only those data very near ($\sim$100 km distance or less) plate boundaries and those data likely reflecting unmodeled processes. Thus data from continental margins have been excluded on the basis of possible contributions from sediment loading or thermal contraction - effects not modeled, and data are not used from regions of complex tectonics not likely to be a response solely to large-scale forces (e.g., Alps, Appalachians, and North America west of the Rockies).

Based on a comparison to the observed stress orientations in Figure 1, the predicted stresses are in best agreement with the observations when pushing forces at ridges are included in the driving force model and when the net pulling force due to subducted lithosphere is comparable in magnitude or is at most a few times larger than other forces acting on the plates. On the basis of intraplate stresses, therefore, resistive forces opposing the motion of the slab with respect to the mantle must nearly balance the negative buoyancy of the relatively cool, dense slab, in agreement with similar conclusions derived from other considerations (Smith and Toksöz, 1972; Forsyth and Uyeda, 1975, Richter, 1977). The maximum ratio of net slab pull to net ridge push is not sensitive to a rate dependence for net slab pull or to the inclusion of other forces in the system. Forces resisting further convergence at continental collision zones along the Eurasian plate are important for intraplate stresses, and improve the fit to the data in Europe, Asia, and the Indian plate. Resistive viscous drag forces acting on the base of the plate in a direction opposite to "absolute" plate velocity improve the fit to the intraplate stress field for several plates (e.g., Nazca, South America). The intraplate stress field is not very sensitive to an increased drag coefficient beneath old oceanic lithosphere compared to young oceanic or continental lithosphere. Increasing the drag coefficient beneath continents by a factor of five or ten changes the calculated stresses only slightly and has little effect on the overall fit of calculated stresses to observed stresses as long as some resistive drag acts beneath oceanic lithosphere.

Models in which drag forces drive (i.e., act parallel to "absolute" plate velocity) rather than resist plate motions are in poor agreement with the data. This poor agreement may depend on the oversimplified model of the adopted interaction between the plate and the asthenosphere. As noted above, the actual flow pattern in the mantle, including counterflow and possible multiple scales of convection, may be considerably more complicated than has been assumed in these models.

Two models that provide reasonably good fits to a large fraction of the intraplate stress orientation data are shown in Figures 2 and 3. In Figure 2 are shown the predicted intraplate stresses for a model with the following forces: (i) a symmetric pushing force at ridges equivalent to a compressive stress of 100 bars across a 100 km thick plate, (ii) a symmetric pulling force at trenches of the same absolute magnitude, (iii) a symmetric resistive force at continental collision zones of the same absolute magnitude and (iv) a drag stress $-D\underline{v}$, where $\underline{v}$ is absolute plate velocity in cm/yr and D is 0.1 bar/cm/yr beneath oceans and 0.6 bar/cm/yr beneath continents. Note that only the relative magnitudes of these forces are constrained by the stress orientation data; their absolute magnitudes are uncertain to within a multiplicative constant.

The predicted directions of principal stresses for this force model are in good agreement with the data for eastern North America, Europe, Asia near the Himalayas, and the Indian plate. The fit to the data is good in South America, especially away from the trench, and in western Africa and is acceptable in most of the Pacific plate. The orientation of the calculated maximum compressive stress in the Nazca plate for the model is only in moderate agreement with the orientation inferred from the single fault plane solution available. The fit to the data in the northern Pacific, eastern Asia, and east Africa is rather poor. The fit to the data in the northern Pacific and eastern Asia could probably be improved if subduction zone or drag forces were decreased along the western Pacific plate margin or if slab forces were concentrated on the subducted plate. No attempt, however, has been made to vary plate boundary forces locally to match inferred stresses. If such an approach were adopted, most observed stresses could probably be matched but the solution for the driving mechanism would be unjustifiably arbitrary and non-unique.

In Figure 3 are shown the intraplate stresses for a force model that takes the approach of Davies (1978) and Richardson

(1978) based on the assumption that drag balances the net torque on each plate due to boundary forces. The resulting drag thus varies from plate to plate and need not bear a simple relationship to relative plate motions, in contrast to the drag derived from absolute plate motion models consistent with known relative velocities (Solomon and Sleep, 1974; Solomon et al., 1975; Minster et al., 1974). The force model includes: (i) a symmetric force at ridges equivalent to a compressive stress of 100 bars across a 100-km thick plate, (ii) a symmetric force at continental convergence zones of twice this magnitude, (iii) a pulling force at trenches, on the subducted plate only, equivalent to an extensional stress of 100 bars across a 100-km thick plate, and (v) a viscous drag on each plate, due to the rotation of the plate with respect to the underlying mantle (which may be moving), that is necessary to balance the torque on each plate from the boundary forces.

The predicted stress directions for this model (Figure 3) agree very well with the data for several areas. In the North American and Nazca plates, the orientation of the maximum compressive stress is well matched by the model. The fit is almost as good in Europe and in Asia north of the Himalayas. In the Indian plate, compressive stresses trend NW-SE in continental India, in agreement with the data, but the fit is poorer in Australia. In South America, the maximum compressive stress trends E-W, in only moderate agreement with the data. In the Pacific and the eastern part of the African plate the agreement with the data is poor. On the whole, the model provides a better fit to continental than oceanic data, and suggests that any force pulling the overthrust plate toward the trench is probably lower in magnitude than the net pull on the subducted plate.

Discussion: From the standpoint of deviatoric stress magnitudes, the most important general conclusion from the modeling of plate tectonic stresses and the comparison with intraplate stress orientation data is that ridge pushing forces are an important element of the set of driving forces for the models that provide the best fit to observations. The stresses that arise from ridge topography are 200-300 bars compression, as noted above. We are thus led to the conclusion that regional deviatoric stresses in plate interiors are of this same general magnitude, or 200-300 bars to within a factor of perhaps 2 to 3.

This conclusion should be tempered, however, by several general observations on the results of the plate tectonic stress models. The models represented in Figures 2 and 3, though providing good matches to the data for a number of regions of well characterized stresses, do not fit all of the data. Thus either there are simple models not tested that provide a better fit to the global data set than those shown, or the stress observations are influenced by processes not included in the simple models. Even if a model were obtained that fit all reliable observations to within their estimated errors, it is likely on the basis of models tested to date that this model would not be unique. Thus statements based on elements of best-fitting force models must be made in cognizance of this nonuniqueness.

Stress Magnitudes and Local vs. Regional Stresses

An alternative approach to constrain the magnitude of regional deviatoric stresses in the lithosphere from stress orientation data and plate tectonic models is to find situations in which observed stress orientations are sensitive in approximately equal measure to a local stress field that may be readily quantified and to a regional stress field whose magnitude is to be determined. Such an approach holds high promise for constraining the magnitudes of plate tectonic stresses in oceanic lithosphere.

Consider the effect of a volcanic load on oceanic lithosphere. Such a load leads to lithospheric flexure and to potentially large local bending stresses. For a very large load, such as Hawaii, the local stresses may be in excess of 1 kbar (Walcott, 1970; Watts and Cochran, 1974) and may dominate the regional stress. That bending stresses may dominate regional stresses for Hawaii is supported by the report by Rogers and Endo (1977) that greatest compressive stress axes from composite fault plane solutions for many mantle earthquakes beneath and near the island of Hawaii are radial with respect to the island.

For loads lesser in magnitude than Hawaii, there is the strong prospect that the local stresses are comparable in magnitude to the regional stresses. Thus the mechanisms of earthquakes in the vicinity of such loads might be expected to indicate P and T axes which differ somewhat from regional trends but which are not predictable simply from stress models for the local load only. For earthquakes near very small loads or distant from any pronounced topographic relief, the mechanisms should reflect the regional stress field.

As an illustration of this approach, consider the region near the Ninetyeast ridge in the central Indian Ocean. Th Ninetyeast ridge is a pronounced linear feature some 5000 long and rising 1500-2000 m above the surrounding seafloor (e.g., Bowin, 1973). The ridge is isostatically compensated except at short wavelengths (Bowin, 1973; Detrick and Watts, 1979). Several large earthquakes have occurred in the Indian plate in the general vicinity during this century (Sykes, 1970; Stein and Okal, 1978).

The orientation of principal stresses in the Indian plate may be estimated from the fault plane solutions of intraplate earthquakes. Figure 4 shows the P axis orientations for several large earthquakes in the Indian plate near the Ninetyeast ridge. There is a strongly regionally consistent NW-SE trend to the direction of inferred greatest compressive stress.

Two aspects of this general consistency are noteworthy: (i) The P axes for strike-slip events on and near the Ninetyeast ridge trend in general agreement with those for thrust events in the plate off the ridge. Thus while a zone of weakness associated with the ridge may control the type of faulting (Stein and Okal, 1978), the inferred direction of maximum horizontal stress for Ninetyeast ridge events is still reliable. The data in Figure 4 are entirely consistent with a generally uniform stress field across the portion of the Indian plate shown, with strike-slip rather than thrust motion the preferred fault type within weak zones in the lithosphere. (ii) The P axes for thrust events off the Ninetyeast ridge are not orthogonal to the strike of the ridge. Thus stresses associated with ridge topography do not dominate the local stress field.

This second conclusion can be quantified to produce a constraint on the magnitude of the regional stress field. Adopting Bowin's (1973) model for the isostatic compensation of the Ninetyeast ridge, the compressive force that the ridge exerts per unit length on the lithosphere beneath the adjacent abyssal plain may be estimated from equations (47-49) in Artyushkov (1973):

$$\begin{aligned}\Sigma &= \int (\sigma_{xx} - \sigma_{zz})\,dz \\ &= -\,(.14\zeta + .067\zeta^2) \times 10^9 \text{ bar-cm} + \Sigma_{ridge}\end{aligned} \qquad (1)$$

where σ_{xx} and σ_{zz} are horizontal and vertical normal stress components ($\sim$ principal stresses), ζ is the height of the ridge (in km) with respect to the abyssal plain, Σ_{ridge} is the value of Σ beneath the ridge, the integral is taken over the depth range of horizontal density variations, and the minus sign denotes a compressive force. Note that (1) includes the effects of topography and isostatic compensation only; the effects of viscous forces at the base of the plate and of thermal stress due to any differential cooling between the ridge and surrounding sea floor, for instance, are not included. For ζ = 1.5 to 2 km (Bowin, 19[illegible]), (1) gives $\Sigma - \Sigma_{ridge}$ =

- (0.35 to 0.54) x 10^9 bar-cm, or the equivalent of 70 to 110 bars additional horizontal deviatoric stress over a 50 km thick plate. For comparison, Artyushkov (1973) gives - 1.2 x 10^9 bar-cm and 240 bars compression for the force/length and stress associated with spreading ridges.

Thus the regional deviatoric stresses in the Indian plate (excluding the contribution from the Ninetyeast ridge) must be larger than ~100 bar in magnitude in order to account for the pattern of stress orientations in Figure 4. This lower bound does not, of course, offer guidance as to what the magnitudes of regional deviatoric stresses are, but the result is at least consistent with the inference made above that regional stresses are similar in magnitude to the stresses produced by ridge forces, which are 3 ± 1 times as large as the force exerted by Ninetyeast ridge topography.

A number of other oceanic intraplate earthquakes large enough so that their focal mechanisms are known have occurred in close proximity to prominent bathymetric features (Figure 5). Several of these features involve lithospheric loads that should lead to bending stresses larger than the stresses indicated above for the Ninetyeast ridge. Thus it may be possible by a combination of detailed stress models and careful source mechanisms to bound regional deviatoric stress magnitudes from both above and below using this approach.

Two potential difficulties with this approach should, however, be noted: (i) Many oceanic intraplate earthquakes occur in or near such obvious zones of weakness as fracture zones and volcanic areas (Figure 5). Stress directions inferred from earthquake mechanisms for such events should be used only with caution in the absence of corroborative information from events removed from the weak zone (e.g., Figure 4). (ii) Bending stresses associated with lithospheric flexure are extremely sensitive to depth. Thus for an observation of stress directions from an earthquake mechanism to be a useful constraint on stress amplitude, the focal depth must be known with high precision, probably to within a few kilometers.

Conclusions

The global data on directions of principal stresses in plate interiors can serve as a test of possible plate tectonic force models. Such tests conducted to date favor force models in which ridge pushing forces play a significant role. For such models, the general magnitude of regional deviatoric stresses is comparable to the 200-300 bars compressive stress exerted by spreading ridges.

An alternative approach to estimating magnitudes of regional deviatoric stresses from stress orientations is to look for regions of local stress either demonstrably smaller than or larger than the regional stresses. The regional stresses in oceanic intraplate regions are larger than the $\sim$100 bar compression exerted by the Ninetyeast ridge and less than the bending stresses ($\geq$1 kbar) beneath Hawaii.

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Figure Captions

Figure 1. A summary of intraplate stress orientation data (Richardson et al., 1979). Filled circles denote fault plane solutions; arrows denote P and T axes, where nearly horizontal. Filled circles without arrows denote thrust faults with poorly constrained P axes. Open circles represent in situ data; the line gives the direction of maximum horizontal compressive stress.

Figure 2. Principal horizontal deviatoric stresses in the lithosphere for a model of plate driving forces (see text). Principal stress axes without arrows and with arrows pointing outward denote deviatoric compression and tension, respectively. Relative magnitude of principal stresses is indicated by the length of stress axes. From Richardson et al. (1979).

Figure 3. Principal horizontal deviatoric stresses in the lithosphere for an alternative driving force model in which basal shear balances the torque due to boundary fouces for each plate (see text). From Richardson et al. (1979).

Figure 4. Summary of focal mechanism solutions for the Ninetyeast ridge region of the Indian plate. Lines through filled circles denote the orientations of P Axes. The isolated filled and open circles denote poorly constrained thrust and normal fault solutions. Data from Sykes (1970), Fitch (1972), Sykes and Sbar (1974), Stein and Okal (1978) and Bergman and Solomon (1979).

Figure 5. Histograms of oceanic intraplate earthquakes sorted by likelihood of association with either intraplate zones of

weakness or with local sources of stress (Bergman and Solomon, 1979).

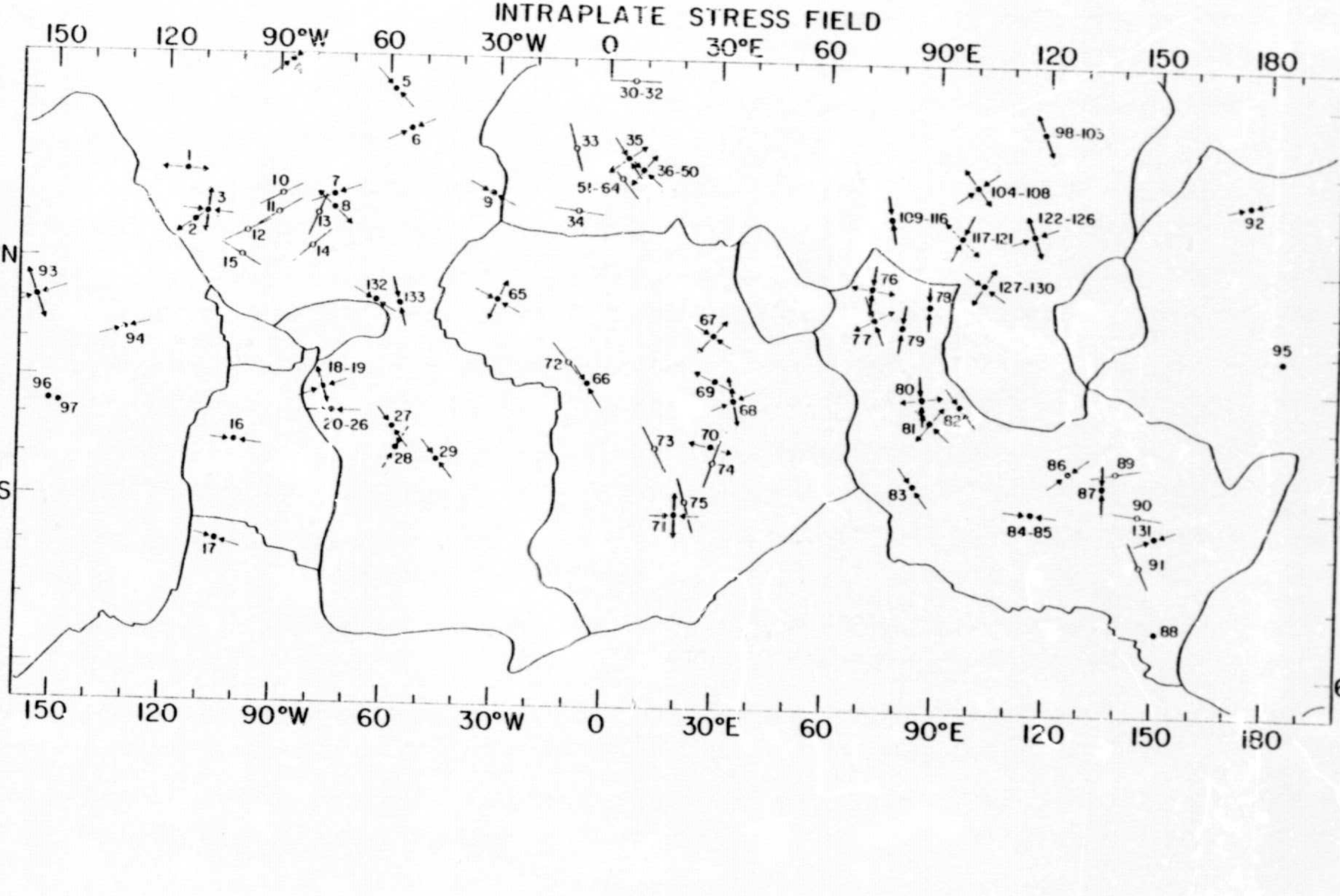
INTRAPLATE STRESS FIELD
150
120
90°W
60
30°W
0
30°E
60
90°E
120
150
180
60
30°N
0
30°S
60

Figure 1

MODEL E 31

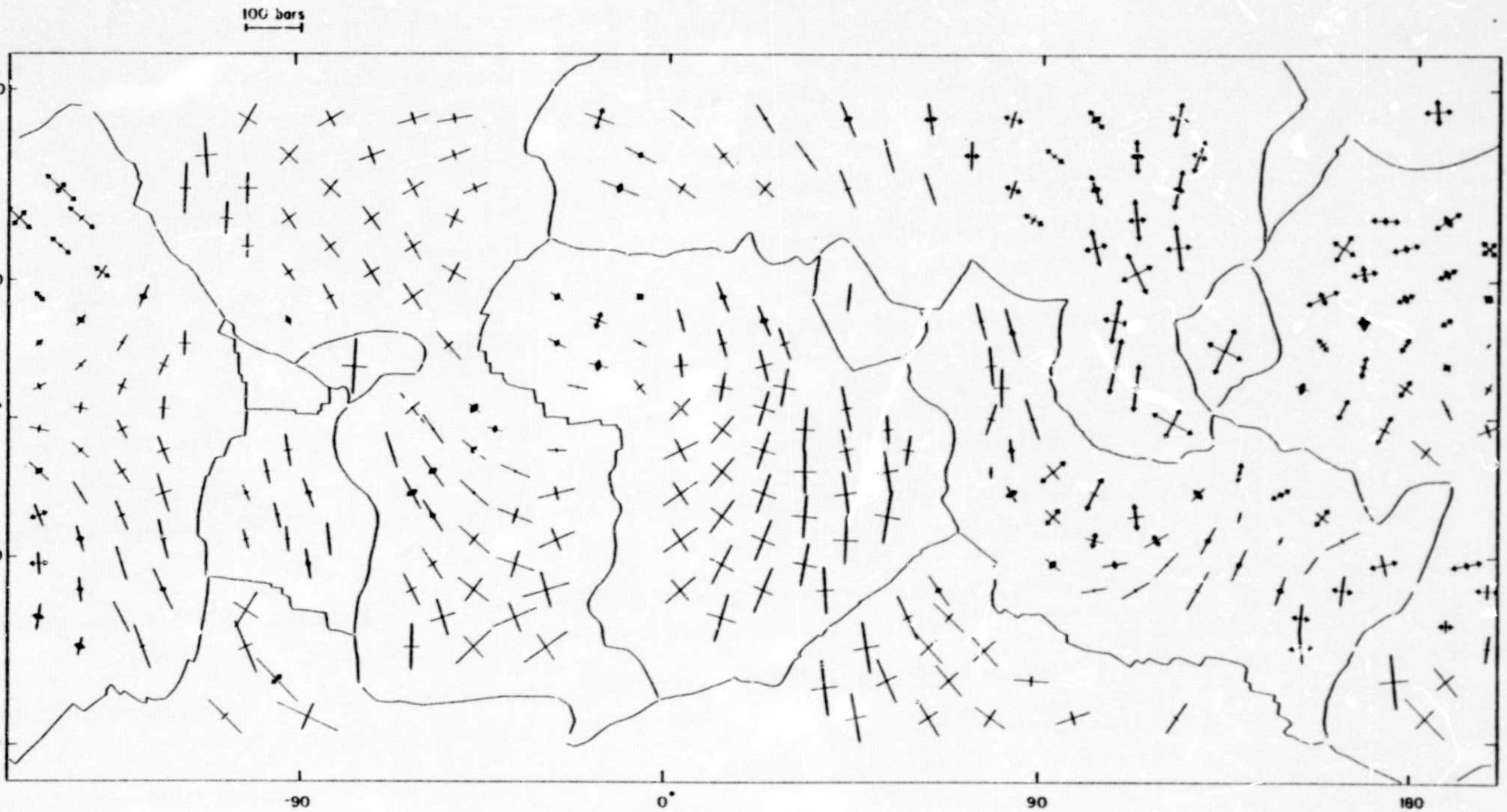

Figure 2

MODEL E29

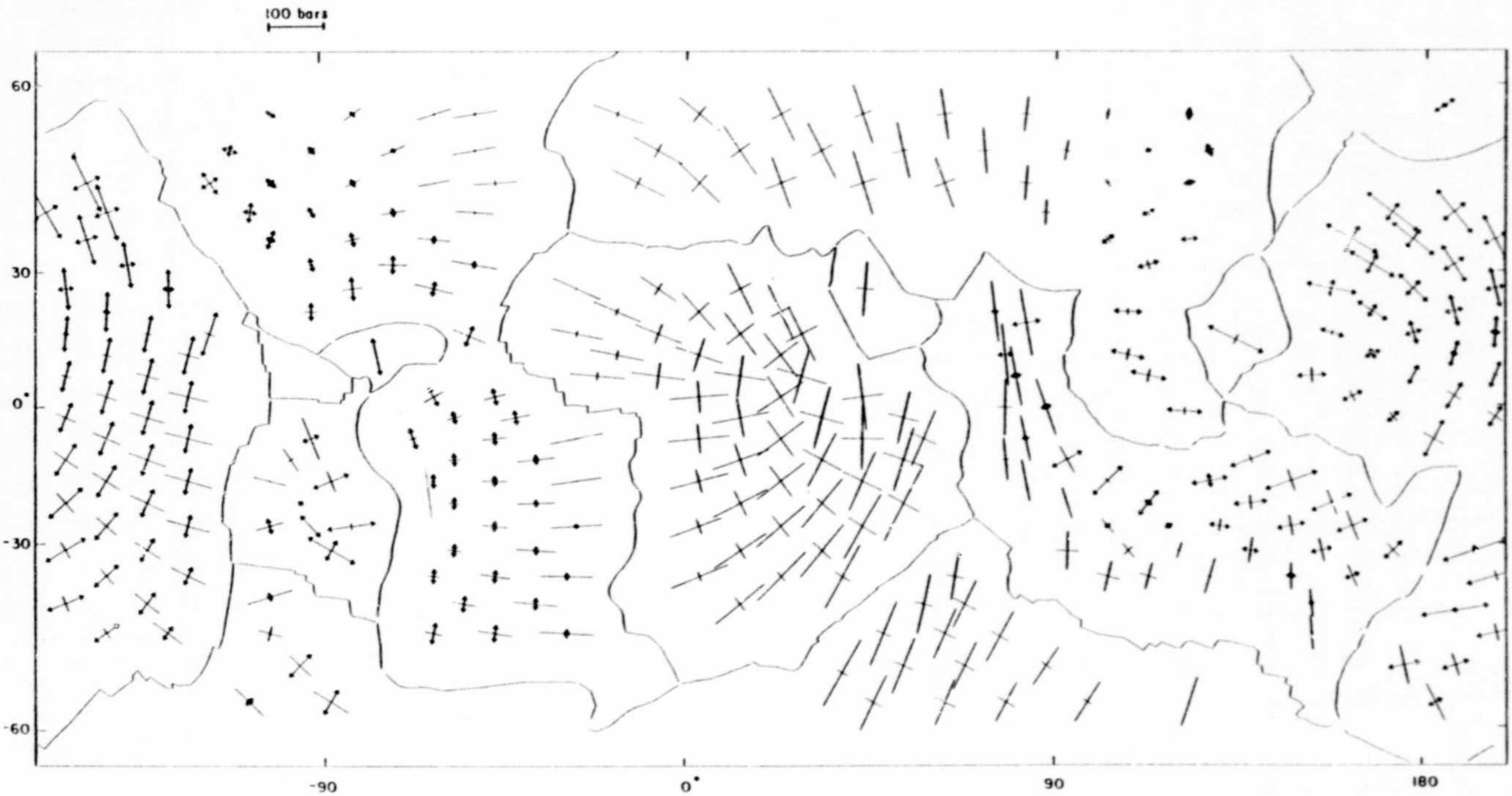

Figure 3

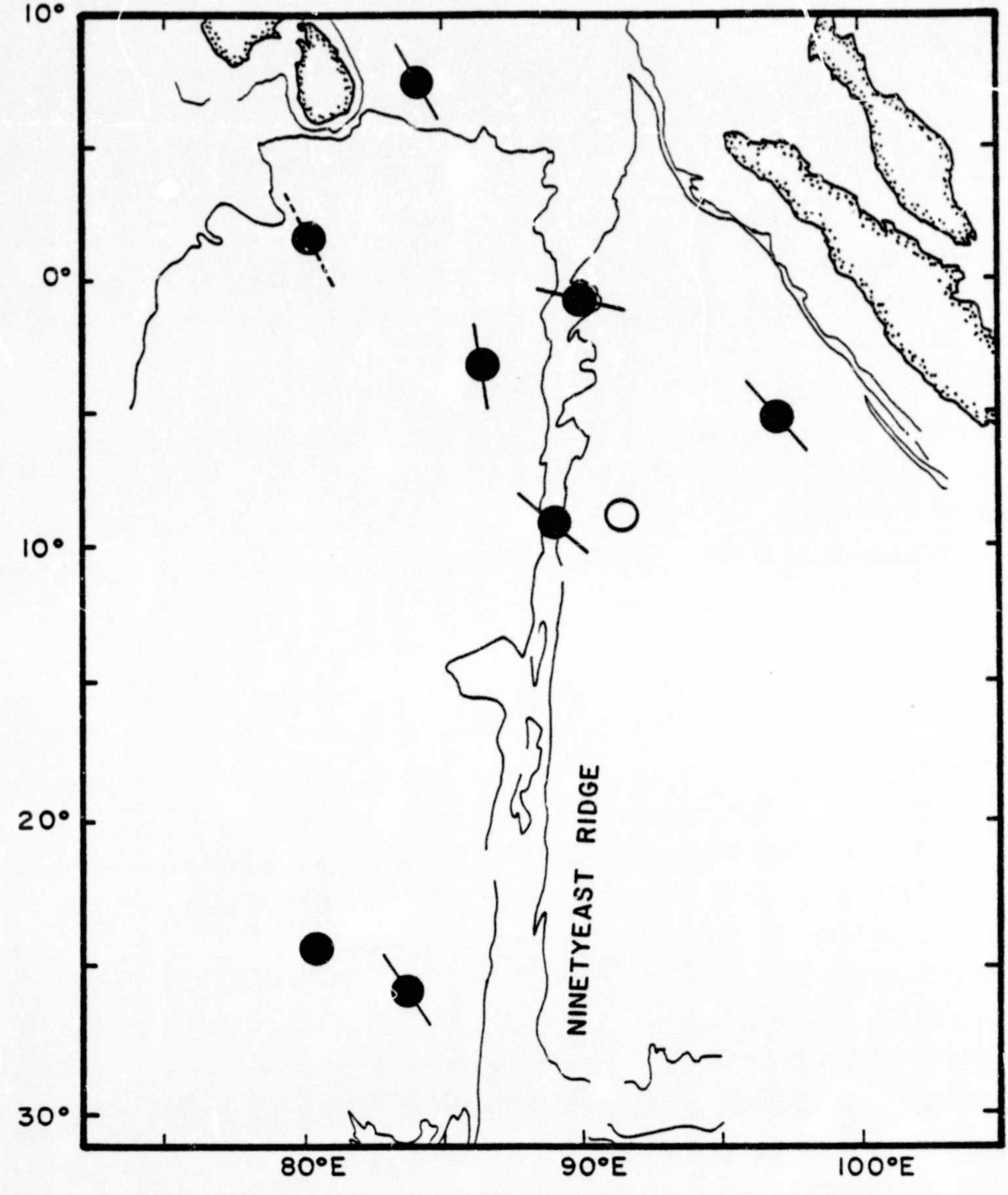

Figure 4

ASSOCIATION WITH ZONES OF WEAKNESS
AND SOURCES OF LOCAL STRESS