**MECHANICAL PROPERTIES AND MODELING
OF SEAL-FORMING LITHOLOGIES**

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TABLE OF CONTENTS

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DOE FORM 1332.16	i
SUMMARY SHEET	2
Research Goals and Accomplishments	2
Continued Research Support	5
TECHNICAL PROGRESS REPORT	6
Project Objectives	6
Publications.....	6
Salt Tectonics Modeling.....	8
Mechanical Properties of Shale	13
CONTINUED RESEARCH	18
REFERENCES.....	19
APPENDIX A	
APPENDIX B	

SUMMARY SHEET

Mechanical Properties and Modeling of Seal-Forming Lithologies DE-FG87ER13711

Research Goals and Accomplishments

Significant progress has been made in this third and final year of funding towards evaluating the roles of rocksalt and shale in the development of structural traps to hydrocarbons. Our research accomplishments summarized here are inclusive over the full three year period and include references to two publications which resulted from previously funded DOE research but which appeared in print during this project period. Research completed over this project period includes three manuscripts that have been reviewed and accepted for publication, and one paper in print (see Appendix A). Support for this research has contributed to the completion of two first rate Ph.D dissertations and an excellent M.S. thesis. W.T. Shea defended and graduated (December, 1991) during the first year of support and is currently employed at Exxon Production Research. R. Mazariegos recently defended his Ph.D. dissertation on the mechanical modeling of mobile salt structures (see Appendix B) and graduated this year (May, 1993). He plans to continue his modeling efforts over the next two years, working with us as a postdoctoral research fellow. W. Ibanez has completed his experimental research of shale deformation, a portion of which appears in a manuscript which has been reviewed and accepted for publication (see publications list of Technical Progress Report) and intends to defend his M.S. thesis shortly.

The general goals of our research supported by DOE Basic Energy Sciences have been to determine the mechanical properties of two important sedimentary lithologies, rocksalt and phyllosilicate-bearing shales, to understand the physics of their deformation, and to follow the development of large scale structures as governed by their mechanical properties. Rocksalt and shale formations are observed to deform under depositional and tectonic loads and share similar roles in the development of structural traps to hydrocarbons. Constitutive relationships for these rock types have been determined and constrained by performing laboratory deformation experiments over widely varying conditions, particularly in strain rate. Numerical modeling of salt structures has proceeded using laboratory-based constitutive relationships for rocksalt and comparing the model predictions with stratigraphic, microstructural, and other geologic constraints for known salt structures.

Specific goals and accomplishments of this research include:

1) the evaluation of models of salt diapir ascent that involve either power law, dislocation creep as determined experimentally by *Horseman et al.* (1993) or linear, fluid-assisted creep as reported by *Spiers et al.* (1988,1990, 1992). We have compared models assuming these two, experimentally evaluated flow laws and examined the predictions they make regarding diapir incubation periods, ascent velocities, deviatoric stresses and strain rates.

2) the evaluation of the effects of differential loading on the initiation and evolution of salt structures. Differential loading conditions are a natural consequence of deltaic depositional cycles and our model results indicate that gradients in load are extremely important to the geometries of salt structures as well as their spacing and rates of growth.

3) examination of the role of basement faults on the initiation and morphologic evolution of salt structures. Continental margins such as the Gulf of Mexico have had an extensional tectonic history and the basement rocks below evaporite sequences are likely to contain numerous normal faults. Again, our models indicate that perturbations at the lower boundaries of mobile salt units are important and significant horizontal flow within salt structures may result..

4) evaluation of the mechanical properties of shale as a function of pressure and determination of the nature of its brittle-ductile transition. Most previous studies of shale deformation have been performed at conditions that promote brittle mechanisms of deformation. Our study has revealed a broad range of conditions over which transitional brittle-ductile deformation is important.

5) evaluation of the mechanical anisotropies of shales with varying concentrations, distributions and preferred orientations of clay. Measurements of clay fabrics in a wide selection of shales and our experiments of the Wilcox formation shale indicate that shales are not very anisotropic in their inelastic mechanical properties.

6) the determination of temperature and rate dependencies of strength for a shale constitutive model that can be used in numerical models that depend on viscous formulations. We have determined parameters that describe the thermally activated deformation of shale in the absence of transient pore pressures and are ready to incorporate them into our finite-volume models of salt tectonics to represent the behavior of sediments overlying salt bodies.

7) determination of the mechanisms of deformation for argillaceous rocks over a wide range of conditions. Our electron and optical microscopy reveal both dilatant, brittle mechanisms of deformation as well as dislocation glide and kinking in clays.

8) evaluation of the effects of H₂O within clay interlayers, as adsorbed surface layers. Our deformation experiments and IR spectroscopy suggests that there may be two mechanisms by which H₂O weakens shales, one through reductions in effective confining pressures, and the second through changes in strength associated with interlayer bonding in clays.

Continued Research Support

Over the remaining months of this project period, we plan to complete and submit for publication our results on salt modeling and write an additional manuscript on effects of clay interlayer H₂O on shale deformation. Our research objectives over the next three years appear in our renewal proposal (report no. DOE/ER 13711-A-3) submitted to DOE BES in February, 1993.

TECHNICAL PROGRESS REPORT

Project Objectives

Both rocksalt and shale lithologies are actively deformed in response to gravitational loads associated with deposition, and both form effective seals to petroleum and natural gas. Thus, the shapes of seal-forming units, and the nature of fractures and faults that may breach them depend upon either the mechanical properties of salt or those of shale, and the loading histories to which they have been subjected.

The objectives of this research include the determination of the mechanical properties under controlled conditions of well-characterized shales, and the numerical modeling of rocksalt (and eventually shale) formations using laboratory-based rheologies. Our numerical modeling efforts to simulate the development of salt structures and our experimental efforts to determine a rate-dependent deformation law for shale have proceeded somewhat independently as well-determined rheologies for rocksalt were available at the outset of the project that could be incorporated into our models, whereas well-constrained rate-dependent constitutive relationships for shales were unavailable. However, both studies have been carried out with a common goal of examining their roles in the development of structural traps. Further research that involves greater integration of our experimental and modeling studies has been proposed in a renewal grant for the next three year period. In the following, we list publications that have appeared in print or have been accepted for publications within this project period, followed by brief summaries of our work on salt tectonics modeling and shale deformation. Appendices A and B include respectively a recent reprint on the unique, very low strain rate experiments (10^{-9}s^{-1}) performed on rocksalt (run durations of 3.5 and 3.6 years) and the recently defended Ph.D. dissertation of Dr. Ruben Mazariegos.

Publications

Doctoral Theses

Shea, W.T. (1991) *The roles of Fabric and Micas in the Deformation of Foliated Rocks*, Doctoral Dissertation, Texas A&M University, 172 pp.

Mazariegos, R.A. (1993) *Mechanical Modeling of the Growth of Salt Structures*. Doctoral Dissertation, Texas A&M University, 217 pp. (see Appendix B).

Abstracts

Shea, W.T., and Kronenberg, A.K. (1991) Stable shear in experimentally deformed mica-rich rocks, *EOS Trans. AGU*, 72, 449.

Mazariegos, R.A. and Russell, J.E. (1992) Growth of diapiric salt structures: integration of experimental and numerical work, *EOS Trans. AGU*, 73, 572.

Ibanez, W.D., Kronenberg, A.K. and Wenk, H.-R. (1992) Anisotropy and the brittle-ductile deformation of shale, *EOS Trans. AGU*, 73, 309.

Referred Papers

Gottschalk, R.R., Kronenberg, A.K., Russell, J.E. and Handin, J. (1990) Mechanical anisotropy of gneiss: failure criterion and textural sources of directional behavior, *J. Geophys. Res.* 95, 21613-21634.

Shea, W.T. and Kronenberg, A.K. (1992) Rheology and deformation mechanisms of an isotropic mica schist, *J. Geophys. Res.* 97, 15201-15237.

Horseman, S.T., Russell, J.E., Handin, J. and Carter, N.L. (1993) Slow experimental deformation of Avery Island salt, *Seventh International Symposium I*, Elsevier Publ. Co., Amsterdam., 64-74 (see Appendix A).

Carter, N.L. Horseman, S.T., Russell, J.E. and Handin, J. (1993) Rheology of rocksalt, *J. Struct. Geol.* (in press).

Shea, W.T. and Kronenberg, A.K. (1993) Strength and anisotropy of foliated rocks with varied mica contents, *J. Struct. Geol.* (in press)

Ibanez, W.D. and Kronenberg, A.K. (1993) Experimental deformation of shale: mechanical properties and microstructural indicators of mechanisms, *Proc .U.S. Symp. Rock Mech. 34th*, (in press).

Figure 1. Velocity vector field evolution for the power law and fluid assisted law models. Both models use and initial amplitude $A_0=100$ m and $\lambda=13.5$ km.

Salt Tectonics Modeling

We have completed the implementation of a 2D finite-volume numerical model to simulate the evolution of salt structures. The model is based on conservation laws for physical systems, a marker equation to keep track of the salt/sediment interface and two laboratory-based rheological laws for rocksalt. The momentum, the continuity, the energy and marker equations are solved by way of a fully-conservative finite-volume technique; by covering the field of interest with finite volume cells, the governing equations of the flow field are reduced to a set of coupled linear algebraic equations. Solution of these equations is implemented by means of a SIMPLE type algorithm (*Patankar and Spalding, 1972*), using iterations of pressure and velocity corrections to satisfy the momentum, and continuity equations simultaneously. The constitutive laws compared in this work employ the equivalent stress $\sigma = \sqrt{3 J_2}$ where J_2 is the second invariant of the deviatoric stress tensor. In addition, equivalent viscosity is defined as $\mu_{eq} = \sigma/(3\epsilon)$ (*Griggs, 1939*), where ϵ is the strain rate which is power conjugate to the equivalent stress σ . Because the equivalent salt viscosity depends on the primitive flow variables, the equivalent viscosity was included in the iterative procedure. The model computes the evolution of the salt/sediment interface as well as the evolution of the velocity, equivalent stress, equivalent strain rate, equivalent viscosity and temperature fields at selected elapsed times during evolution after salt deposition.

A two-layer model incorporating a salt/sediment interface, perturbed by an initial amplitude of 100 m and wavelength of 13.5 km, was examined. This model was used to compare the evolutionary pattern of a salt structure when two different rheologic laws for rocksalt are assumed; we compare the low stress, low strain rate power law ($\epsilon \propto \sigma^{3.4}$) of *Horseman et al. (1993)* and the fluid-assisted diffusion law ($\epsilon \propto \sigma/Td^3$) of *Spiers et al. (1990)*, where T is the temperature in K and d the grain size in mm. The salt and sediment thicknesses are 2.5 and 2.3 km respectively, the density contrast between the two materials is 0.1 g/cm³, sediment viscosity $\mu_2=3.0 \times 10^{19}$ Pa-s and rocksalt viscosity is calculated iteratively using the equivalent stress and equivalent viscosity.

Model results for power-law creep and the fluid-assisted diffusion law for different stages of salt dome evolution are compared in Figure 1. These plots indicate that the geometries of interfaces predicted by the power-law creep and linear, fluid-assisted creep models are similar, but the power-law model does not attain the same structural maturity as reached by the fluid-assisted diffusion law model for the same time

of evolution. This observation is summarized in Figure 2 in which amplitude of the perturbing anomaly (A) and interface particle velocity (measured at the midpoint of the model) are shown as functions of elapsed time (My). The rapid evolution of structures predicted by our models using the fluid-assisted diffusion law is a direct result of the weakening effect of water on the shear strength of rocksalt (*Urai et al., 1986; Spiers et al., 1988*). From our numerical experiments the fluid-assisted diffusion law predicts evolution of salt structures 2.5 times faster than the power-law creep model.

The evolution of the equivalent stress field (in MPa) has been examined as well and typical results for the power-law creep and the fluid-assisted diffusion law models are shown in Figure 3. The equivalent stress at the top of the salt structure for the power-law creep model (at times varying from 16 to 23 My) remains nearly constant at approximately 0.4 MPa. The lowest values of differential stress inferred from measurements of subgrain boundary spacings in domal rocksalt samples (*Carter et al., 1993*) are somewhat greater (0.6 MPa) than our model results, and microstructural indicators of stress generally indicate larger stresses than those of our model. This discrepancy may be due to our assumption that the sedimentary overburden behaves as a highly viscous fluid having a constant viscosity ($\mu_2 = 3.0 \times 10^{19}$ Pa-s), or to the two-dimensional geometry of the model, or some other assumption that we have made. Comparison of equivalent stresses within the salt for the fluid-assisted diffusion and the power-law creep models at approximately the same structural maturity (e.g. 7.86 My and 19.56 My, respectively) indicates that stresses predicted by the power-law model are at least five times greater than those predicted by the diffusion creep model. At advanced levels of structural maturity (e.g. 11.79 My and 23.59 My, respectively), equivalent stresses predicted by the power-law model may be as much as 100 times greater than those predicted by the diffusion creep model. Thus, when water is available during deformation of rocksalt, differential stresses as small as 0.004 MPa may be sufficient to accelerate the rates of flow.

The evolution of the equivalent salt viscosity field for the power-law creep model depends on velocity gradients, temperature and the experimentally derived power-law creep parameters. Determinations of equivalent viscosity were therefore made within the velocity-pressure iteration. For the power-law model, values of equivalent salt viscosity develop local maxima (Figure 4), reaching values (of up to 1.0×10^{14} MPa-s,) that exceed the viscosity 3.0×10^{13} MPa-s assumed for the sediments. In contrast, the fluid-assisted law (e.g. $\dot{\epsilon} \propto \sigma / Td^3$) viscosity is independent of the equivalent stress (σ), although it is strongly dependent on temperature (T) and grain size (d). For comparison purposes, we assumed a constant grain size of $d = 7.5$ mm (the mean grain size of Avery

Figure 2. Amplitude (m) and interface particle velocity (m/My) versus time (My) for the power law and fluid assisted diffusion models of figure 1. (a) Amplitude (m) vs. time (My). (b) Interface particle velocity (m/My) vs. time (My).

Figure 3. Equivalent stress field evolution for the power law and fluid assisted law models. If water is available during deformation, differential stresses as small as 0.004 Mpa are enough to make rocksalt flow at accelerated rates.

Figure 4. Equivalent salt viscosity field evolution for the power law and the fluid assisted law models. Fluid assisted model equivalent salt viscosity values are lower than those predicted by the power law model by 2 to 3 orders of magnitude depending on the evolutionary stage of the salt structure.

Island rocksalt) and determined equivalent viscosities for the fluid-assisted law (Figure 4). Contour plots indicate that viscosity for this model varies primarily with depth (e.g. decreasing with depth as temperature increases). Viscosities predicted by the fluid-assisted model are lower than those predicted by the power-law creep model by 2 to 3 orders of magnitude depending on the evolutionary stage of the salt structure.

Predictions of strain rate fields at similar, mature stages of development are nearly the same for the two salt rheologies assumed (Figure 5) with dominant values of the order of $4 \times 10^{-15} \text{ s}^{-1}$. These results are in excellent agreement with strain rates (of $3.5 \times 10^{-15} \text{ s}^{-1}$) reported by *Jackson and Talbot* (1986) for salt structures in the Gulf Coast during the pillow stages and more advanced levels of structural maturity. *Seni and Jackson* (1984) calculated strain rates from changes in the relief of salt diapirs over different time intervals in the East Texas Basin; their results are also of the same order of magnitude as the strain rates predicted by our models.

In addition to these simple two-layer models, we have evaluated the effects of deposition during the time of salt dome growth. For this purpose we modeled four loading conditions according to Table 1, for both the power-law creep model (Figure 6) and the fluid-assisted diffusion model (Figure 7).

Table 1
Loading conditions for time-dependent deposition models

Loading Condition Model	Overburden Thickness (km)	Time interval (My)
a	1.2	$0 < t \leq 14$
	3.6	$14 < t \leq 20$
b	1.2	$0 < t \leq 10$
	3.6	$10 < t \leq 20$
c	1.2	$0 < t \leq 6$
	3.6	$6 < t \leq 20$
d	3.6	$0 < t \leq 20$

Figure 5. Equivalent strain rate (s^{-1}) field evolution for the power law and fluid assisted law models. Strain rate numerical results are in good agreement with calculated strain rates for typical Gulf Coast salt structures made by Jackson and Talbot (1986) and with strain rates calculated by Seni and Jackson (1984) for the East Texas Basin.

Figure 6. Power law model for time dependent deposition ($A_0=100$ m and $\lambda=13.5$ km). The salt thickness is $h_{\text{salt}}=1.2$ km and the sedimentary layers thicknesses $L1=1.2$ km and $L2=2.4$ km. Times on the right side of the figure refer to elapsed time after salt deposition. Loading conditions are according to Table 1.

Figure 7. Fluid assisted law model for time dependent deposition ($A_0=100$ m and $\lambda=13.5$ km). The salt thickness is $h_{salt}=1.2$ km and the sedimentary layers thicknesses $l_1=1.2$ km and $l_2=2.4$ km. Times on the right side of the figure refer to elapsed time after salt deposition. Loading conditions are according to Table 1.

For these models, we assumed a constant salt layer thickness of 1.2 km. A sedimentary layer L1 was deposited first (L1=1.2 km) and at a prescribed time a second sedimentary layer L2 was deposited (L2=2.4 km) as indicated by the arrows in Figures 6 and 7. These models employed a density contrast of $\Delta\rho=0.1 \text{ g/cm}^3$, a viscosity for the sediments of $\mu_2=3.0 \times 10^{19} \text{ Pa-s}$, and an initial amplitude of the perturbing anomaly $A_0=100 \text{ m}$. Comparison of Figures 6a and 7a, indicates that the perturbing anomaly for the power-law model (Figure 6a) did not evolve significantly during the elapsed time of 20 My, even after deposition of layer L2 ($t > 14 \text{ My}$). In contrast results for the fluid-assisted diffusion model (Figure 7a) show significant evolution as early as 6 My, before layer L2 is even deposited. Once deposition of layer L2 occurs ($t > 14 \text{ My}$), the salt structure continues to grow until a mature diapiric stage is reached at 20 My. Figures 6b and 6c also show that the power-law model did not evolve significantly even when additional loading (deposition of layer L2) started earlier. In contrast, the fluid-assisted diffusion law model predicts detachment of the salt structure at times of 18 My (Figure 7b) and 14 My (Figure 7c). The power-law model predicts development of mature salt structures in an elapsed time of 20 My (Figure 6d) only when full loading conditions are present (e.g. deposition of layer L1+L2) at the outset. Given this same (and probably unreasonable) loading condition, the fluid-assisted model reaches similar maturity in only 10 My after salt deposition (Fig. 7d) and detachment of the salt structure is observed by 12 My.

Although we have obtained many of the geometries observed for natural domal salt structures, including lateral spreading and detachment of salt, we still need to incorporate constitutive relations for the sediments that more accurately describe the mechanical behavior of the overburden. Our next step will be to use the experimentally derived expression for shale deformation, fitting the nonlinear time-dependent response of shale with a power-law (with a large value of $n \approx 80$). Previous mechanical models that have examined the response of highly nonlinear power-law materials (with large values of n) have shown that deformation is strongly localized within narrow shear zones, assuming only minor initial perturbations in boundary conditions (*Kilsdonk, 1989; Davies, 1990*). We expect that the large values of n for shale applied to our model sediment layers may allow us to observe the development of faulting which is usually associated with the evolution of salt structures. For example, observations of large allochthonous salt sheets in the Alaminos Canyon region, northwest Gulf of Mexico (*Hardin, 1939*) have led to the suggestion that faulting may be induced below allochthonous salt sheets due to deformation of the infilling sediments as the stem feeder thins and detaches (Figure 8). Indeed, these faults may provide possible hydrocarbon

Figure 8. Simplified diagram showing sequential development stages leading to the presence of shallow, detached salt structures. Compare to detached salt structures of figures 7b, 7c and 7d (After Hardin, 1989).

migration pathways. In summary, results of our time dependent deposition models indicate that the morphology of salt structures changes depending on loading conditions, the times of loading, and the rheology of rocksalt.

We have evaluated the effects of differential loading on the initiation and evolution of salt structures using sedimentation rates (e.g. loading conditions) of the northwest East Texas Basin and the low stress, low strain rate power-law creep ($\epsilon \propto \sigma^{3.4}$) of *Horseman et al.* (1993). For the East Texas Basin salt tectonics setting, two general observations have been made by *Harris and McGowen* (1987); first, the structural maturity seems to be related to the original thickness of salt, and second, the salt apparently migrated at different times in different parts of the East Texas Basin by several mechanisms, including faulting, gravity sliding (caused by basinward tilting) and mass imbalance by uneven loading.

The differential loading conditions applied and corresponding model results are shown in Figures 9 and 10 respectively. An initial salt layer thickness of 1 km was assumed and two progradational cycles were imposed. The first depositional cycle (over times of $0 < t \leq 5$ My) represents deposition of the Gilmer formation (TG) (Figure 9a), with thicker sediments deposited on the left side of the figure. As a result of this differential loading, a small perturbing anomaly may be observed at $t = 5$ My (Figure 10). The second progradation cycle, over the time interval $5 < t \leq 15$ My, represents deposition of the Travis Peak formation (TTP) and was modeled by the addition of sediments and advance of the step in sediment thickness (Figure 9b). Lateral flow of salt modeled for this interval occurs from left to right and has formed a salt ridge which has reached the surface by 10 My. This salt ridge is predicted to spread laterally from 10 to 15 My and salt pillow structures start to develop at both sides of the ridge by 15 My (Figure 10). By 15 My, the Travis Peak formation was assumed to have advanced across the model and differential sediment loading ceased (Figure 9c); however, perturbing anomalies have already been induced. During the time interval $15 < t \leq 26$ My, the salt ridge is no longer predicted to spread laterally, since it is no longer exposed to the surface, and advances vertically instead (Figure 10). By 26 My, pillow salt structures at both sides of the salt ridge are predicted to have evolved to mature domal stages. The last depositional event modeled represents the uniform deposition of the First Anhydrite Stringer (FAS) during the time interval $26 < t \leq 35$ My (Figure 9d), and model results are marked by evolution of the three salt structures to more mature stages (Figure 10).

The effect of differential loading on the initiation and further development of salt and shale structures has long been recognized and *Kehle* (1971) suggests that differential loading is the dominant mechanism responsible for salt dome initiation. Differential

Figure 9. Differential loading conditions induced by two progradational cycles, the first one took place for $0 < t < 5$ My and the second one for $5 < t < 15$ My. Northwest East Texas Basin sedimentation rates are used. TG=Top of Gilmer, TTP=Top Travis Peak and FAS=First Anhydrite Stringer.

Figure 10. Numerical modeling of differential loading conditions of figure 9. The perturbing anomalies are induced by the effect of two progradational cycles which occurred for the time intervals $0 < t < 5$ My and $5 < t < 15$ My. Times on the right of the figure refer to elapsed time after salt deposition. TG=Top of Gilmer, TTP=Top Travis Peak and FAS=First Anhydrite Stringer.

loading may be particularly important if power-law creep governs the flow of rocksalt since viscosities are sensitive to differential stress and therefore pressure gradients within the model (*Mazariegos*, 1993). Differential loading has been inferred to be important in the northwestern part of the East Texas Basin during the Late Jurassic and Early Cretaceous (*Harris and McGowen*, 1987), and salt flow appears to have occurred laterally away from regions of thickened sediments associated with a withdrawal basin. Our numerical results (Figure 10) predict the same patterns of salt structure evolution as postulated by *Harris and McGowen* (1987), when low stress, low strain rate power-law creep is assumed.

Currently we are evaluating the effects of basement faulting on the genesis and evolution of salt structures; at this stage we have modeled a two-layer system for different fault geometries using Newtonian rheologies for rocksalt and sediments with viscosities of $\mu_1 = 3.0 \times 10^{17}$ Pa-s and $\mu_2 = 3.0 \times 10^{19}$ Pa-s, respectively. Figure 11 shows the evolution of salt structures for salt wedges and underlying basement faults of varying dimensions, as may be present at basin margins of extensional tectonic settings. Diapirs are rooted at the basement discontinuities and the morphology of the evolving diapirs depends on the type and size of discontinuity, and the amount of salt available. In extensional tectonic settings (pull-apart basins) several basement discontinuities may be present, and Figure 12 shows model results for salt structures developed over two normal faults. The morphologies of salt diapirs modeled are distinctly asymmetric by 40 My, especially the diapir on the right side of the figure in which salt flow occurred preferentially to the right. This type of structure resembles salt nappe structures of the Gulf of Mexico such as the Sigsbee Escarpment (Figure 13). It is important to point out that our current basement faulting models are based on linear rheologies, and the time scale should not be interpreted strictly. We plan to continue our modeling studies of basement faulting including time-dependent deposition, differential loading and nonlinear rheologies for both salt and sediments.

Mechanical Properties of Shales

We have nearly completed a suite of triaxial experiments on an illite-bearing shale from the Wilcox Formation, Louisiana designed to investigate its strength as a function of confining pressure (P_c), temperature (T), strain rate ($\dot{\epsilon}$) and orientation relative to bedding with molecular water restricted to clay interlayers, (i.e., in the absence of a free pore fluid phase). We have used a wide variety of methods to characterize the composition and texture of the starting material and have performed deformation experiments following a

Figure 11. Salt wedge plus normal basement fault models. The models use linear rheologies for both salt and sediments. Salt viscosity $\mu_1 = 3.0 \times 10^{17}$ Pa.s, sediment viscosity $\mu_2 = 3.0 \times 10^{19}$ Pa.s and density contrast $\Delta\rho = 0.1$ g/cm³. The morphology of the evolving diapirs depend on the type and size of the basement discontinuity and the amount of salt available.

Figure 12. Two dipping normal fault model. The dipping normal faults induce perturbing anomalies which evolve to large salt diapirs rooted at basement discontinuities. The horizontal and vertical dimensions are 28.6 km and 2.4 km respectively. Linear rheologies are used in the model.

Figure 13. Interpreted north-south seismic profile across a salt wedge of the central Sigsbee Bulge. TCDM- top of the Cinco de Mayo seismic unit (uppermost Miocene to lower Pliocene); TUMR- top Upper Mexican Ridges (middle Miocene to upper Miocene); TLMR- top Lower Mexican Ridges (lower Eocene to middle Oligocene); TCP- top Campeche (middle Cretaceous to lower Eocene); MCSB- middle Cretaceous sequence boundary. (After Simmons, 1992).

strategy similar to that used in our previous studies of mica-bearing rocks. The data provide a means of comparing the mechanical responses of clay-bearing shales with mica-bearing schists. Transmission electron microscopy (TEM) of the deformed shale samples indicates that deformation mechanisms include brittle microcracking, frictional sliding, dislocation glide, and kinking of clays, much as we have found for mica-bearing rocks.

Shales are generally difficult to characterize and they encompass a wide range of fissile fine-grained rocks with varied compositions and textures. A critical portion of this project has been the detailed characterization of the starting material by a combination of X-ray diffraction (XRD) and other standard methods of soils analysis, infrared (IR) absorption, X-ray pole figure goniometry, scanning electron microscopy (SEM), and high resolution TEM. Based upon weighted averages of powder XRD patterns (e.g., Figure 14) for disaggregated and sized fractions of the starting material, measurements of cation exchange, and shifts in clay reflections upon heat treatment and glyceration (*Gibbs, 1965; Jackson, 1956; Klug and Alexander, 1954; Kunze, 1965; Mehru and Jackson, 1960*), the shale starting material is made up of illite (40-50%), kaolinite (5-10%), quartz (30-40%), feldspar (1-5%), and chlorite (1-5%) with small concentrations of smectites and/or mixed layer clays. The cementing minerals appear to be primarily quartz and aragonite.

IR measurements were made using powder samples of the starting material embedded in dry KBr. Absorption spectra (Figure 15) reveal both hydroxyl and molecular water in the clays including sharp OH-stretching bands assigned to hydroxyl in illite and kaolinite and broad bands at 3400 cm^{-1} and 1625 cm^{-1} assigned to the OH stretch and HOH bend modes, respectively, of molecular water in illite and smectite interlayers (*van Olphen and Fripiat, 1979*). The total water content of the shale starting material (2.1 wt %) was determined by measuring the weight loss of a disaggregated sample following heat treatment at 220°C for 72 hours.

The preferred orientation of illite grains within the starting material has been determined by X-ray goniometry using the (001) reflection. Illite basal planes are preferentially oriented parallel to bedding as indicated by broad maxima of (001) poles perpendicular to bedding (Figure 16), with maximum X-ray intensities of ~ 4 times the mean density expected for a random orientation distribution. Maxima of this order appear to be characteristic of a wide number of shales we have sampled, and are much smaller than are typically observed in higher grade slates and phyllites.

The starting material has been examined by optical microscopy, SEM, and TEM. High resolution TEM lattice imaging combined with electron diffraction was used to identify individual clay grains (Figure 17) and look for any interlayering of mixed clays.

Figure 14. Powder X-ray diffraction patterns of coarse clay fraction of Wilcox formation shale, following K and Mg exchange treatments, heat treatment, and glyceration.

Figure 15. Powder IR absorbance spectrum of Wilcox formation shale at wavenumbers exhibiting O-H stretching bands of illite and kaolinite associated with structural hydroxyl groups (OH) and molecular water (H₂O) within illite interlayers.

Illite 001 Preferred Orientation

Contour Interval = 0.50

Minimum = 0.00

Maximum = 4.16

Figure 16. Contoured stereograph of the illite (001) reflection intensity for Wilcox formation shale, showing broad maxima perpendicular to bedding (contour interval is in multiples of the intensity expected for a random distribution of (001) orientations).

a

Figure 17. High resolution TEM micrographs of shale starting material from the Wilcox formation. (a) Illite platelets with wavy and discontinuous (001) layers (with unit cell spacing of 1.01 nm). (b) Chlorite grain with parallel, well formed (001) layers 4-11 nm in dimension (vertical boundary is a low angle tilt wall)

b

Figure 18 TEM micrograph of illite platelets of the shale starting material (Wilcox formation) with widely varying orientations relative to bedding and fine detrital quartz within domain of high porosity with microcracks and polygonal voids.

In addition, TEM of ion-milled samples has revealed complex microstructures associated with deposition and burial down to very small scales. Extremely fine illite grains show complex patterns of growth with open nm-scale voids (Figure 18). Polygonal voids are common, associated with incompatible grain shapes and changes in clay orientation. Regions of particularly high porosity appear to be associated with clay grains that have nearly random crystallographic orientations in regions that locally may have escaped compaction and preserve original depositional textures of clays (*Bennett et al.*, 1981). Basal planes of illite (Figure 17a) are typically wavy, similar to those observed in other shales by *Ahn et al.* (1983) and *Lee et al.* (1985). Direct lattice imaging using high resolution TEM has revealed basal layer spacings of 0.7nm, 1.0nm, and 1.4nm for individual clay grains as well as traces of irregularly layered and highly defective smectites. However, despite pronounced HOH absorption bands in the IR and suggestions of mixed-layer clays in XRD patterns, we have not found significant regions of intergrown mixed-layer clays.

The triaxial, constant strain rate experiments on shale samples of the Wilcox formation are nearly finished with results available over temperatures $21^{\circ} \leq T \leq 200^{\circ}\text{C}$, confining pressures $20 \text{ MPa} \leq P_c \leq 400 \text{ MPa}$, and strain rates $2.3 \times 10^{-7} \text{ s}^{-1} \leq \dot{\epsilon} \leq 1.0 \times 10^{-3} \text{ s}^{-1}$ (Table 2). Several experiments were performed at multiple conditions, step-wise decreasing strain rate $\dot{\epsilon}$ or increasing temperatures T and the results of these experiments are listed in Tables 3 and 4, respectively. Right circular samples were loaded parallel, perpendicular, and at 45° to bedding, and samples loaded perpendicular to bedding were deformed with varying initial water contents. Stress-strain curves (Figure 19a) typically exhibit transients in differential stress to strains of 5-6%, after which stresses are relatively constant with occasional, small stress drops for samples that developed discrete shear fractures. Anisotropy is displayed at all confining pressures tested (Figure 19 a,b) with the highest differential stresses measured perpendicular to bedding, and the lowest stresses measured at 45° to bedding. However, anisotropy of this shale, with its modest clay preferred orientation, is less pronounced than that expressed by higher grade rocks with stronger preferred orientations, such as Martinsburg slate (*Donath*, 1961).

Deformation is brittle at low confining pressures ($P_c \leq 60 \text{ MPa}$) while transitional brittle-ductile behavior is observed at the higher pressures ($P_c \geq 100 \text{ MPa}$), as indicated by sample barrelling accompanied by shear zone development (Table 2). Mohr envelopes for samples of all three orientations are nonlinear (Figure 20) with apparent values of internal friction μ that decrease with increasing normal stress. Samples loaded at 45° to bedding exhibit the greatest curvature and the weakest pressure dependence at high pressures with $\mu \approx 0.3$ (at $\sigma_n \leq 150 \text{ MPa}$) and $\mu \approx 0.02$ (at $\sigma_n = 250 \text{ MPa}$).

Table 2

Experiment Number	Confining Pressure (MPa)	$\dot{\epsilon}$ (s ⁻¹)	T (°C)	($\sigma_1 - \sigma_3$) at yield or plateau (MPa)	$\dot{\epsilon}$ at yield stress (%)	Total $\dot{\epsilon}$ (%)	Fault Plane (° to C axis)	Comments
<i>Load Axis Perpendicular to Bedding</i>								
H 1.1	250	2.3x10 ⁻⁵	22	285.6	6.4	6.6	-	-1.4% H ₂ O
H 1.2	250	2.3x10 ⁻⁵	22	224.0	5.8	7.4	30	0% H ₂ O
H 1.3	250	2.3x10 ⁻⁵	22	294.9	6.5	6.8	-	-1.53% H ₂ O
H 1.5	250	2.3x10 ⁻⁵	19.5	143.5	7.3	8.6	-	+1.1% H ₂ O
H 1.6	250	2.3x10 ⁻⁵	22	317.4	6.0	6.9	-	-2.1% H ₂ O
H 1.8	250	2.3x10 ⁻⁵	22	273.9	6.8	8.2	-	-1.15% H ₂ O
H 1.9	250	2.3x10 ⁻⁵	22	115.2	5.0	6.8	35	+1.27% H ₂ O
H1.10	250	2.3x10 ⁻⁵	21	68.1	4.2	4.2	35	+2.0% H ₂ O
P1.1 D	20	2.2x10 ⁻⁴	22	78.3	1.7	2.8	32	B11 VSR3
P1.2 D	200	2.2x10 ⁻⁴	22	203.0	4.8	6.8	-	B11 S3 VR1
P1.4 D	100	2.2x10 ⁻⁴	21.5	145.8	3.7	10.6	30	shear zone
R1.1 D	250	2.3x10 ⁻⁵	22	221.2	5.9	7.4	-	
R1.2 D	250	2.2x10 ⁻⁴	22	242.6	5.5	8.2	34	
RS1.1 D	250	2.2x10 ⁻⁴	22	209.4	5.4	8.1	-	RS [†]
S1 R1	60	1.0x10 ⁻³	22	147.9	3.5	6.1	37	
S2 R1	120	1.0x10 ⁻³	22	191.9	4.5	12.1	40 & 57	CS [‡]
S3 R4	200	1.0x10 ⁻³	22	249.2	7.9	13.9	21 & 17	CS
T1.3 D	250	2.3x10 ⁻⁵	200	182.1	5.7	6.4	-	
TS1.1 D	250	2.3x10 ⁻⁵	22	219.2	5.9	9.6	-	TS ^Ω
<i>Load Axis Parallel to Bedding</i>								
P2.1 D	50	2.2x10 ⁻⁴	22	113.8	1.1	1.8	30	
P2.2 D	250	2.2x10 ⁻⁴	25	236.6	5.1	5.8	15 & 41	CS - Kinked
R2.1 D	250	2.3x10 ⁻⁵	22	229.2	4.2	8.8	47 & 20	Kink and Fracture
RS2.1 D	250	2.2x10 ⁻⁴	22	218.5	5.2	10.0	-	RS
S1 R2	60	1.0x10 ⁻³	22	148.6	1.9	2.9	21-35	Arced shear
S2 R2	120	1.0x10 ⁻³	22	202.1	3.6	5.7	45 & 36	CS - Kinked
S3 R2	200	1.0x10 ⁻³	22	264.3	6.3	12.6	45 & 40	CS - Kinked
T2.1 D	250	2.3x10 ⁻⁵	200	162.1	3.0	3.4	45	
T2.2 D	250	2.3x10 ⁻⁵	130	175.0	4.3	4.6	0	
TS2.2 D	250	2.3x10 ⁻⁵	22	222.9	3.8	6.0	0	TS
<i>Load Axis at 45° to Bedding</i>								
P 3.1 D	20	2.2x10 ⁻⁴	22	63.1	.9	.9	33	
P 3.2 D	80	2.2x10 ⁻⁴	22	118.1	1.9	2.1	34	
P 3.3 D	140	2.2x10 ⁻⁴	22	140.1	4.0	4.7	-	
P 3.4 D	200	2.2x10 ⁻⁴	21	205	4.7	7.1	-	
P3.101 D	400	2.3x10 ⁻⁵	28	205.2	9.3	10.5	-	Riecker Rig
RS3.3 D	250	2.2x10 ⁻⁴	22	198.3	5.5	8.2	-	RS
RS3.4 D	250	2.2x10 ⁻⁴	22	188.2	6.5	9.5	-	RS
S1 R3	60	1.0x10 ⁻³	22	113.5	2.2	3.2	45	
S2 R3	120	1.0x10 ⁻³	22	167.3	3.3	9.9	34	
S3 R3	200	1.0x10 ⁻³	22	200.5	12.7	12.8	35	
T3.2 D	250	2.3x10 ⁻⁵	129	192.5	6.8	9.7	-	
T3.3 D	250	2.3x10 ⁻⁵	22	204.7	4.9	5.2	-	
TS3.1 D	250	2.3x10 ⁻⁵	22	207.0	5.6	13.6	-	TS
<i>Load Axis at 64° to Bedding</i>								
P 64°	250	2.2x10 ⁻⁴	22	218.3	6.4	8.1	-	
‡	Conjugate Shear							
†	Rate Stepping-Experiment							
Ω	Temperature-Stepping Experiment							

Table 3

Rate-Stepping Experiments

Experiment #	Sample Orientation	Confining Pressure (MPa)	$\dot{\epsilon}$ (s^{-1})	T ($^{\circ}C$)	$\sigma_1 - \sigma_3$ at plateau (MPa)	ϵ at plateau stress (%)	Total ϵ (%)	α (MPa $^{-1}$)	Comments
RS1.1 D	LA* $\perp$ to Bedding	250	2.2×10^{-4}	22	209.4	5.4	8.1	0.36 ± 0.04	$\dagger 10^{-6}$ strain rate run stopped mid-way and restarted 12 hrs. later allowing for 2 stresses
			2.3×10^{-5}		203.0	0.6			
			2.2×10^{-6}		196.9 / 194.1 $\dagger$	0.3 / 1.3			
			2.3×10^{-7}		190.8	0.09			
RS2.1 D	LA // to Bedding	250	2.2×10^{-4}	22	218.5	5.2	10.0	0.30 ± 0.04	Fracture formed at end of 10^{-7} step
			2.3×10^{-5}		214.1	0.7			
			2.2×10^{-6}		205.8	1.4			
			2.3×10^{-7}		195.4	1.4			
RS3.3 D	LA 45° to Bedding	250	2.2×10^{-4}	22	198.3	5.5	8.2	0.50 ± 0.06	Fracture formed at end of 10^{-7} step
			2.3×10^{-5}		193.1	0.5			
			2.2×10^{-6}		187.4	0.8			
			2.3×10^{-7}		185.0	0.5			
RS3.4 D	LA 45° to Bedding	250	2.2×10^{-4}	22	188.2	6.5	9.5	0.50 ± 0.04	Fracture formed at end of 10^{-7} step
			2.3×10^{-5}		184.5	0.9			
			2.2×10^{-6}		180.2	1.0			
			2.3×10^{-7}		174.4	1.0			

Table 4

Temperature-Stepping Experiments

Experiment #	Sample Orientation	Confining Pressure (MPa)	$\dot{\epsilon}$ (s^{-1})	T ($^{\circ}C$)	$\sigma_1 - \sigma_3$ at plateau (MPa)	ϵ at plateau stress (%)	Total ϵ (%)	Q (kJ/mol)	Comments
TS1.1 D	LA* $\perp$ to Bedding	250	2.3×10^{-5}	22	219.2	5.9	9.6	87 $\pm$ 1	Q obtained by combination of runs TS1.1 and T1.3
TS2.2 D	LA // to Bedding	250	2.3×10^{-5}	70	205.9	3.0	6.0	124 $\pm$ 9	Q obtained by combination of runs TS2.1, T2.1 & T2.2
TS3.1 D	LA at 45° to Bedding	250	2.3×10^{-5}	70	203.2	5.3	13.6	181 $\pm$ 11	
				22	207.0	5.6			
				73	189.9	7.9			
				133	167.2	10.5			
				196	154.2	13.3			

* Load Axis

Figure 19. Results of constant strain rate experiments performed on shale samples at $T = 22^\circ\text{C}$, $P_c = 250$ MPa, and $\dot{\epsilon} = 10^{-4} \text{ s}^{-1}$. (a) Differential stress - strain data for samples loaded parallel, perpendicular, and at 45° to bedding. (b) Differential stress (taken after the yield point once stress is relatively constant with increasing strain) as a function of loading direction β relative to bedding. Data for shale are compared with much stronger anisotropy reported for Martinsburg slate (Donath, 1961).

Figure 20. Mohr envelopes for shale samples loaded perpendicular, parallel, and at 45° to bedding ($T = 22^\circ\text{C}$, using data at strain rates over $10^{-4} \geq \dot{\epsilon} \geq 10^{-5} \text{ s}^{-1}$).

Most previous experimental studies of shale deformation have emphasized the roles of brittle deformation mechanisms, from macroscopic fracture to microscopically brittle, macroscopically ductile cataclasis and compaction (*Handin et al.*, 1963; *Chenevert and Gatlin*, 1965; *McLamore and Gray*, 1967; *Braddock and Machette*, 1976; *Van Eeckhout*, 1976; *Peng and Okubo*, 1978; *Chong et al.*, 1979; *Bauer*, 1980; *Smith and Cheatham*, 1980; *Tandanand*, 1985; *Yelverton*, 1985; *Sargand and Hazen*, 1987; *Swan et al.*, 1989; *Steiger and Leung*, 1988, 1990, 1991) while a combination of cataclasis, basal shear and kinking has been reported by *Nuesch* (1989, 1991). In all previous studies, brittle sliding and shear has been assumed to occur by homogenous slip on hydrated clay layers rather than by dislocation glide. TEM observations of our deformed shale samples have confirmed the importance of microcrack propagation as well as dilatancy associated with kink bands within extremely localized shear zones (down to 6 μm in width) (Figure 21). However, in addition to these brittle deformation mechanisms we observe evidence for dislocation glide (Figure 22) and crystal plasticity leading to micro-kink band formation (Figure 23) in samples shortened at confining pressures P_c of 200 to 250 MPa.

Given the TEM evidence for a component of dislocation glide at the higher confining pressures and the reduced sensitivity of strength to confining pressure P_c , we have chosen to fit results of strain-rate-stepping and temperature-stepping experiments at $P_c = 250$ MPa to an exponential glide law of the form

$$\epsilon = A \exp \{ \alpha (\sigma_1 - \sigma_3) \} \exp (-Q/RT) \quad (1)$$

where Q is an activation energy and α and A are material constants. Results of strain-rate-stepping tests (Figure 24a) have been recast as differential stress vs. the natural logarithm of strain rate to determine values of α ($1/\alpha = d(\sigma_1 - \sigma_3)/d \ln \epsilon$ at constant T). Similarly, the results of temperature-stepping experiments have been plotted as differential stress vs. reciprocal temperature (Figure 25) to determine values of Q/α ($Q/\alpha R = d(\sigma_1 - \sigma_3)/d(1/T)$ at constant rate) and, combined with measured values of α , determine activation energies Q .

The Wilcox shale at high confining pressures is semi-brittle in all orientations tested. However, the relative contributions of various deformation mechanisms may differ for samples loaded in different orientations. Correspondingly, the observed temperature and strain rate effects appear to depend on bedding orientation; for samples shortened perpendicular to bedding $Q = 66$ kJ/mol and $\alpha = 0.36$ MPa⁻¹, whereas samples loaded parallel to bedding yield $Q = 145$ kJ/mol and $\alpha = 0.30$ MPa⁻¹. Remarkably, these parameters are similar to those describing the deformation of biotite schist ($Q = 89$

Figure 21. TEM micrograph of shale sample shortened at 45° to bedding (sample S3 R3) within narrow dilatant ($6\text{ }\mu\text{m}$ wide) shear zone (scale bar is $0.5\text{ }\mu\text{m}$).

Figure 22. TEM micrograph of basal dislocations in illite ($g = 020$) within several μm of optical-scale shear zone of sample (S3 R2) shortened parallel to bedding (scale bar is $0.5 \mu\text{m}$).

Figure 23. High resolution TEM micrograph of kinked chlorite (1.41 nm lattice spacing) and illite (1.01 nm lattice spacing) and associated polygonal voids in shale (sample T1.3D) compressed perpendicular to bedding (scale bar is 50 nm).

Figure 24. Results of strain rate-stepping experiment. (a) Stress-strain curve for shale at sequential strain rates of 10^{-4} , 10^{-5} , 10^{-6} , and 10^{-7} s^{-1} . (b) Recast as differential stress, taken after the yield point once stress is relatively constant with strain, versus the natural logarithm of strain rate, the data can be used to determine parameter α .

Figure 25. Results of temperature-stepping experiment performed on shale, plotted as differential stress versus reciprocal temperature.

kJ/mol, $\alpha = 0.15$ to 0.55 MPa^{-1} , *Shea and Kronenberg, 1992*) and (001) dislocation glide in muscovite ($Q = 47 \text{ kJ/mol}$, $\alpha = 0.5 \text{ MPa}^{-1}$, *Mares and Kronenberg, 1993*) and biotite ($Q = 82 \text{ kJ/mol}$, $\alpha = 0.41 \text{ MPa}^{-1}$, *Kronenberg et al., 1990*) single crystals.

Most of our experiments have been conducted on samples with $\sim 2.1 \text{ wt\%}$ water, presumably within clay interlayers as characterized by the broad O-H stretching band at 3400 cm^{-1} (Figure 15). However, we have performed additional experiments on samples with variable amounts of water, from 0 to $4.1 \text{ wt\%}$ by driving interlayer water off prior to deformation by heat treatment ($T = 80\text{--}200^\circ\text{C}$) in air or by adding water by exposure to conditions of high relative humidity (98%) at temperatures of $\sim 60^\circ\text{C}$ (using a controlled humidity chamber). Differential thermal analyses of illite at these temperatures (*van Olphen and Fripiat, 1979*) indicate that interlayer H_2O is readily removed by heat-treatment in air at these temperatures and the measured increases in weight of our samples held in the controlled humidity chamber suggest that interlayer water may be adsorbed within relatively short times (~ 96 hours). We have yet to carefully examine O-H absorption bands in illite powder samples taken from the deformed samples. However, the initial results, using weight changes as a measure of water content, indicate that water has a significant effect on the strength of shale (Figure 26). We plan to complete our current investigation of shale deformation by disaggregating deformed samples and preparing IR powder mounts under controlled conditions and studying changes in the broad 3400 cm^{-1} absorption band associated with interlayer molecular water.

The results of our studies of shale deformation have several important implications for the constitutive relations of argillaceous rocks. The common assumption made in modeling studies that argillaceous rocks exhibit perfect plasticity (e.g., *Chapple, 1978*; *Davis et al., 1983*; *Stockmal, 1983*; *Dahlen, 1984*) is largely confirmed. However, shales deformed in the absence of pore fluids show intrinsic temperature and rate effects that are small (but measurable) and consistent with the onset of dislocation glide at high confining pressures. We have fit our high pressure results to an exponential glide law. However, for purposes of modeling, the data may be used to fit the parameters of a power-law with a large value of n (~ 80). Thus, we plan to make significant improvements in our modeling studies, incorporating a more realistic rheology for the sediments (largely made up of shale) surrounding growing salt structures.

The transition from brittle to more nearly ductile deformation of shale is very broad and a wide range of dilatant as well as volume-conserving mechanisms continue to operate at the highest pressures ($P_c = 400 \text{ MPa}$) and temperatures ($T = 200^\circ\text{C}$) investigated. The mechanisms that give rise to semi-brittle deformation include collapse

Figure 26. Results of constant strain rate experiments performed on heat-treated and hydrated shale samples at $T = 22\text{ }^{\circ}\text{C}$, $P_c = 250\text{ MPa}$, $\dot{\epsilon} = 10^{-5}\text{ s}^{-1}$. The initial water content of the starting material is 2.1 wt%.

of polygonal voids, microcrack extension, frictional sliding, dislocation glide on (001) clay interlayers, and micro-kink formation. The extent to which each of these mechanisms contributes to deformation of shales at geologic strain rates may differ from their relative contributions to deformation in our experiments. However, given that the mean stresses that shales are subjected to in sedimentary basins are small relative to those applied in our experiments, we expect that natural deformation of shales will generally involve semi-brittle, dilatant mechanisms.

An additional contribution of these studies to our understanding of shales involves the role of water in deformation. Our results suggest that interlayer surface layers of water influence strength. At present, we do not know whether this form of water influences microcracking or glide since the processes of opening (001) cleavage and dislocation motion on (001) both involve breaking bonds of the illite interlayers. However, interlayer water is thought to have properties that differ from those of pore fluids, and its effect on mechanical properties (*Homola et al.*, 1989; *Van Alsten and Granick*, 1989; *Thompson and Robbins*, 1990) may differ from pore fluids that influence effective stress.

CONTINUED RESEARCH

Within the remaining three months of this project period, we expect to complete and submit two manuscripts on our numerical results on salt tectonics, derived from R. Mazariegos' thesis research (Appendix B). We are continuing IR measurements of water within clay interlayers of heat-treated and hydrated shale samples with the goal of refining our understanding of H₂O weakening effects. We anticipate that these results may lead to an additional publication on shale deformation. At the end of this project period, we are hopeful that funding under our renewal proposal will commence and that we can proceed to incorporate the non-linear time-dependent mechanical expression for shale in our numerical models. In addition, we plan to examine the time-dependent deformation of fluid-saturated shales over strain rates that will promote fluid pressure perturbations leading to dilatancy hardening/softening and capture the transition to mechanical response that can be described macroscopically by an effective-pressure law.

REFERENCES

- Ahn, J.H., Lee, J.H., and Peacor, D.R. (1983) Mineralogical and textural transitions in phyllosilicates during burial diagenesis of Gulf Coast shales, *Geol. Soc. Amer. Abstr. with Prog.* 15, 512.
- Bauer, R.A. (1980) The loss of natural moisture content of Pennsylvanian shale and effect on physical properties, *Proc. Illinois Mining Institute , 87th Annual Meeting*, Illinois Institute of Natural Resources, State Geological Survey Division, 13-28.
- Braddock, W.A. and Machette, M. (1976) *Experimental Deformation of Pierre Shale*, U.S. Air Force Geophysics Laboratory Report #AFGL-TR-76-0086, U.S. Air Force, Hanscom AFB, MA, 96 pp.
- Carter, N. L. and Hansen, F. D. (1983) Review -Creep of rocksalt, *Tectonophysics*, 92, 275-333.
- Chenevert, M.E., and Gatlin, C. (1965) Mechanical Anisotropies of laminated sedimentary rocks, *J. Soc. Petrol. Engrs.*, 5, 67-77.
- Chapple, W.M. (1978) Mechanics of thin-skinned fold-and-thrust belts, *Geol. Soc Amer., Bull.*, 89, 1189-1198.
- Chong, K.P., Uenishi, K. and Munari, A.C. (1979) Three dimensional characterization of the mechanical properties of Colorado oil shale, in: *Proc. of the 20th U.S. Sympos. on Rock Mechanics*, eds. K.E. Gray, Univ. of Texas at Austin, 369-379.
- Dahlen, F.A., Suppe, J., and Davis, D.M. (1984) Mechanics of fold-and-thrust belts and accretionary wedges: cohesive Coulomb theory, *Jour. Geophys. Res*, 89, 10087-10101.
- Davies, R.K. (1990) *Models and Observations of Faulting and Folding in Layered Rock*, Ph.D. dissertation, Texas A&M University, College Station, Texas, 147 pp.

- Davis, D.M., Suppe, J., and Dahlen, F.A. (1983) Mechanics of fold-and -thrust belts and accretionary wedges, *Jour. Geophys. Res.*, 88, 1153-1172.
- Donath, F. (1961) Experimental study of shear failure in anisotropic rocks, *Geol. Soc. Am. Bull.*, 72, 985-990.
- Gibbs, R.J. (1965) Error due to segregation in quantitative clay mineral X-ray diffraction mounting techniques, *Amer. Min.*, 50, 741-751.
- Griggs, D.(1939), Creep of rocks, *Journal of Geology*, 47, 225-251.
- Handin, J., Hager, R.V. Jr., Friedman, M., and Feather, J.N. (1963) Experimental deformation of sedimentary rocks under confining pressure: pore pressure tests, *Amer. Assoc. Petrol. Geol. Bull*, 47, 717-755.
- Hardin, N. S. (1989) *Allochthonous Salt Sheets in the Alaminos Canyon Region, Northwest Gulf of Mexico: Present Configuration and Processes of Emplacement*, Ph.D. Dissertation, Texas A&M University, College Station, Texas, 202 pp.
- Harris, D. W., and McGowen, M. K., (1987), Influence of differential sediment loading on salt tectonics in the East Texas Basin, in: *Dynamical Geology of Salt and Related Structures*, (eds) I. Lerche and J. J. O'Brien, Academic Press, Boston.
- Homola, A.M., Israelachvili, J.N., Gee, M.L., and McGuiggan, P.M. (1989) Measurements of and relation between the adhesion and friction of two surfaces separated by molecularly thin liquid films, *Jour. Tribology*, 111, 675-682.
- Horseman, S. T., Carter, N. L., Russell, J. E. and Handin, J. (1993) Slow experimental deformation of Avery Island salt, *Seventh International Symposium on Salt, 1*, Elsevier Publ. Co., Amsterdam, 67-74.
- Jackson, M.L. (1956) *Soil Chemical Analysis -Advanced Course*, Department of Soils, University of Wisconsin, Madison, WI.
- Jackson, M. P. A., and Talbot, C. J., (1986), External shapes, strain rates, and dynamics of salt structures: *Geol. Soc. America Bull.*, 97, 305-323.

- Kehle, R. O. (1971) *Origin of the Gulf of Mexico*, The University of Texas at Austin, Geology Library, call no. q557K260, unpublished report, unpaginated.
- Kilsdonk, M.W. (1989) *A Study in Extensional Tectonics*, Ph.D. dissertation, Texas A&M University, College Station, Texas, 200 pp.
- Klug, H.P. and Alexander, L.E. (1954) *X-ray Diffraction Procedures for Polycrystalline and Amorphous Materials*, J. Wiley and Sons, New York, 716 pp.
- Kronenberg, A.K., Kirby, A.K., and Pinkston, J. (1990) Basal slip and mechanical anisotropy of biotite, *Jour. Geophys. Res.*, 95, 19257-19278.
- Kunze, G.W. (1965) Pretreatment for mineralogical analysis, in: C.A. Black (editor), *Methods for Soil Analysis- Part I. Agronomy*, 9, ASA, Inc., madison, WI, 568-577.
- Lee, J.H., Ahn, J.H., and Peacor, D.R. (1985) Textures in layered silicates: progressive changes through diagenesis and low-temperature metamorphism, *J. Sed. Pet.*, 55, 532-540.
- Mares, V.M. and Kronenberg, A.K. (1993) Experimental deformation of muscovite, *Jour. Struct. Geol.* (in press).
- Mazariegos, R. A. (1993) *Mechanical modeling of the growth of salt structures*, Ph.D. Dissertation, Texas A&M University, College Station, Texas, 217 pp.
- McLamore, R., and Gray, K.E. (1967) The mechanical behavior of anisotropic sedimentary rocks, *J. Eng. and Ind. Res.* 89, 62-76.
- Mehra, O.P. and Jackson, M.L. (1960) Iron oxide removal from soils and clays by a dithionite-citrate system buffered with sodium bicarbonate, *Clays and Clay Min.*, 7, 317-327.
- Nuesch, R. (1991) *Das Mechanische Verhalten von Opalinuston*, Ph.D. Dissertation #9349, Eidgenössischen Technischen Hochschule (ETH), Zürich, 244 pp.

- Nuesch, R. (1989) Experimental deformation of shale compared with natural features, in: *Deformation Mechanisms, Rheology and Tectonics*, ed. R.J. Knipe and E.H. Rutter, The University of Leeds.
- Patankar, S. V., and Spalding, D. B. (1972) A calculation procedure for heat, mass and momentum transfer in three-dimensional parabolic flows, *Int. J. Heat Mass Transfer*, 15, 1787-1806.
- Peng, S.S. and Okubo, S. (1978) Prediction of in-situ stresses from directional properties of rock cores for field development of Devonian Shales, *EGSP Open File Report #108*, Morgantown Energy Research Center, Morgantown, W. Virginia, 74 pp.
- Sargand, S.M. and Hazen, G.A. (1987) Deformation behavior of shales, *Int. J. Rock Mech. Min. Sci. and Geomech. Abstr.* 24, 365-370.
- Seni, S. J., and Jackson, M. P. A., (1984), *Sedimentary record of cretaceous and tertiary salt movements, East Texas Basin: Times, rates, and volumes of salt flow and their implications for nuclear waste isolation and petroleum exploration*, The University of Texas at Austin, Bureau of Economic Geology, Report of Investigations No. 139.
- Shea, W.T. and Kronenberg, A.K. (1992) Rheology and deformation mechanisms of an isotropic mica schist, *Jour. Geophys. Res.* 97, 15201-15237.
- Simmons, G.R. (1992) *The Regional Distribution of Salt in the Northwestern Gulf of Mexico: Styles of Emplacement and Implications for Early Tectonic History*, Ph.D. Dissertation, Texas A&M University, College Station, Texas, 180 pp.
- Smith, M.B. and Cheatham, J.B. Jr. (1980) A three-dimensional anisotropic yield condition for Green River shale, *J. Energy Res.* 102, 184-189.
- Spiers, C.J., Urai, J.L., and Lister, G.S. (1988) The effect of brine (inherent or added) on rheology and deformation mechanisms in salt rock, in: *Proc. Second Conf. on Mechanical Behavior of Salt* (eds.) H.R. Hardy, Jr. and M. Langer, Trans. Tech. Publ. Hanover, 89-102.

- Spiers, C.J., Schutjens, P.M., Brzesowsky, T.M, Peach, C.J., Liezenberg, J. L.O., and Zwart, H.J. (1990) Experimental determination of constitutive parameters governing creep of rocksalt by pressure solution, in: *Deformation Mechanisms, Rheology and Tectonics*, (eds.) R.J. Knipe and E.H. Rutter, *Geological Society Special Publication No. 54*, 215-227.
- Spiers, C. J. and Brzesowsky, R. H. (1992) Densification behavior of wet granular salt: Theory versus experiment, *Seventh International Symposium on Salt, 1*, (in press) Elsevier Publ. Co., Amsterdam.
- Steiger, R.P. and Leung, P.K. (1988) Quantitative determination of the mechanical properties of shales, in: *Ann. Tech. Conf. Exh. , Oct. 2-6, SPE 18024*, Society of Petroleum Engineers, Houston, TX. 69-78.
- Steiger, R.P. and Leung, P.K. (1990) Predictions of wellbore stability in shale formations at great depth, in: *Rock at Great Depth*, eds. V. Maury and D. Fourmaintraux, A.A. Balkema. , Rotterdam, 1209-1218.
- Steiger, R.P. and Leung, P.K. (1991) Consolidated undrained triaxial test procedure for shales, *Proc. 32nd U.S. Sympos. Rock Mech., July 10-12*, Norman, Oklahoma.
- Stockmal, G.S. (1983) Modeling of large-scale accretionary wedge deformation, *Jour. Geophys. Res.*, 88, 8271-8287.
- Swan , G., Cook, J., Bruce, S., and Meehan, R. (1989) Strain rate effects in Kimmeridge Bay shale, *Int. J. Rock Mech. Min. Sci. and Geomech. Abstr.*, 26, 135-149.
- Tandanand, S. (1985) Moisture adsorption rate and strength degradation of Illinois shales, *26th U.S. Sympos. on Rock Mechanics, June 26-28* , Rapid City, SD, 591-600.
- Thompson, P.A. and Robbins, M.O. (1990) Origin of stick-slip motion in boundary lubrication, *Science*, 250, 792-794.
- Urai, J. L., Spiers, C. J., Hendrik, H. J., and Lister, G. S. (1986) Weakening of rock-salt by water during long-term creep, *Nature*, 324, 554-557.

Van Alsten, J. and Granick, S. (1989) Friction measured with a surface forces apparatus, *Tribology Trans.*, 32, 246-250.

Van Eekhout., E.M. (1976) The mechanisms of strength reduction due to moisture in coal mine shales, *Int. J. Rock Mech. Min. Sci. and Geomech. Abstr.* 13, 61-67.

Van Olphen, H. and J.J. Fripiat (1979) *Data Handbook for Clay Materials and Other Non-metallic Minerals*, Pergamon Press, Oxford, 346 pp.

Yelverton, F.D. (1985) *Experimental Deformation of the Pierre Shale: Room Temperature Tests*, Masters Thesis, University of Colorado, 142 pp.

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